Chapter 3

Neoproterozoic and Cambrian

continental rifting, continent–arc collision
and post-collisional magmatism

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3.1 Introduction

Major advances have been made over the last decade in our understanding of the distribution, compositions, age and significance of Late Neoproterozoic and Cambrian rocks in Victoria. These advances have been driven mainly by the new generation of geological mapping carried out by the Geological Survey of Victoria, using detailed aeromagnetic, gravity and radiometric datasets covering much of the state. In addition, detailed work by universities over certain areas, particularly the Glenelg region in the far west and locally around Stawell, have made significant contributions to the knowledge of the Cambrian geology.

In western Victoria, Cambrian rocks comprise all the known sedimentary and volcanic bedrock as well as numerous granites. Although there is some possibility of Proterozoic rocks occurring here, no dated rocks have returned ages older than Cambrian. The rock units and geological histories of the Glenelg, Grampians–Stavely and Stawell zones, which were poorly known until recently, are now much better understood. In central and eastern Victoria, the Cambrian rocks generally lie at deeper structural levels and are only exposed in the hanging walls of major faults. Important new information is available for the Glenelg River Complex in westernmost Victoria, the exposed and drilled volcanic belts in western Victoria, and the Cambrian greenstones around Pitfield and on Phillip Island.

3.1.1 Proterozoic

There are few rocks in Victoria that can be confidently assigned to the Precambrian. The existence of old, almost entirely hidden, subcrust beneath parts of Victoria has been suggested by previous authors (Scheibner, 1985; Clemens, 1988; Chappell & White, 1992; Cas, 1983; Powell, 1983; Fergusson *et al.,* 1986; Gray *et al.,* 1991; McBride *et al.*, 1996). Its possible presence beneath central Victoria comes from a new interpretation of many geological features and regional magnetic data (VandenBerg *et al.,* 2000; Cayley *et al.,* 2002). In this model, the Melbourne Zone and the eastern part of the Bendigo Zone are underlain by thin Neoproterozoic–Cambrian continental crust — the Selwyn Block — that forms a northern extension of Tasmania. The major definitive characteristic of the Selwyn Block is deformation during the Cambrian, equivalent in time to the Tyennan Orogeny of Tasmania. It has been argued (VandenBerg *et al.,* 2000; Cayley *et al.,* 2002) that slices of the Cambrian cover sequence of the Selwyn Block are exposed as structural windows eroded through the Mount Useful Fault Zone in central eastern Victoria, cropping out as the Licola and Jamieson Volcanics.

3.1.2 Cambrian

Rocks in western Victoria can be assigned to the Delamerian and the Lachlan Fold belts (Fig. 3.1), with the boundary between these presently taken as the Moyston Fault (Figs. 3.2, 3.3) immediately east of the Grampians Ranges (Cayley & Taylor, 1996b, 1998a; Cayley *et al.,* 2002). The Delamerian Fold Belt rocks in western Victoria lie in the Glenelg and Grampians–Stavely zones (Fig. 3.2) (VandenBerg *et al.,* 2000), which have been affected by the 515 to 490-480 Ma Delamerian Orogeny. The Stawell Zone to the east was not deformed until much later, at 450-420 Ma, in an event equivalent in time and effect to the Benambran Orogeny of eastern Victoria (VandenBerg, 1978, 1999; Cayley & McDonald, 1995; Foster *et al.,* 1999).

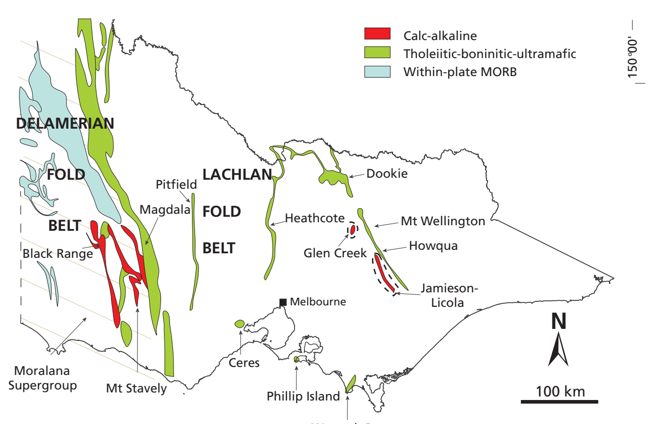
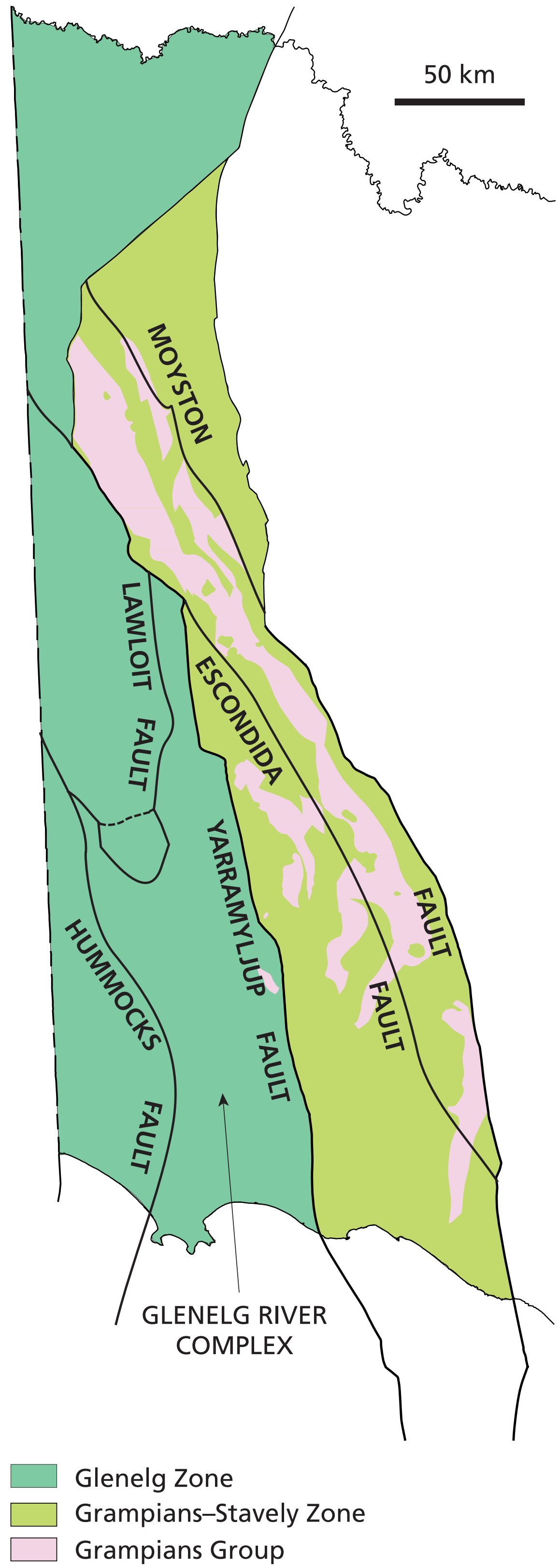


Fig. 3.1: Distribution and affinities of Cambrian and Late Neoproterozoic rocks in the Lachlan and Delamerian Fold belts in Victoria, showing the major greenstone occurrences and inferred extent based on aeromagnetic data (modified from VandenBerg *et al.,* 2000).



3.2 Delamerian Fold Belt

3.2.1 Introduction

The Glenelg and Grampians–Stavely zones are the most westerly geological zones in Victoria and represent the easternmost extension of the Delamerian Fold Belt (Fig. 3.2). The Glenelg Zone consists of the more deformed and higher grade westerly portion including the Glenelg River Complex, whereas the Grampians–Stavely Zone comprises the less deformed and less metamorphosed eastern portion.

In this review, Delamerian Fold Belt rock sequences are described from west to east. Detailed information is available for outcropping Cambrian sequences of the Glenelg Zone exposed in the catchment of the Glenelg River. The Cambrian rocks of the Grampians–Stavely Zone are much more poorly exposed and best known from greenstone-dominated belts outcropping further east in the Black Range, Mount Stavely and Mount Dryden regions. The northward extent and distribution of these and other rocks beneath Murray Basin cover (Fig. 3.1) has been traced using new aeromagnetic and gravity data, with some drillhole control (e.g. Moore, 1996; Maher *et al.,* 1997a).

3.2.2 Glenelg Zone (including the Glenelg River Complex)

The Glenelg Zone comprises the metamorphic and igneous rocks underlying the Dundas Tableland of far western Victoria, including those of the Glenelg River Complex (Wells, 1956; Gibson & Nihill, 1992; Turner *et al.,* 1993; Anderson & Gray, 1994; Kemp & Gray, 1999b; Gray *et al.,* 2002; Kemp *et al.,* 2002). The eastern limit of the Glenelg Zone, and of the igneous–metamorphic complex, is the north-trending Yarramyljup Fault (Fig. 3.3, 3.4), which juxtaposes high-grade metasedimentary rocks against slate and metasiltstone of the Grampians–Stavely Zone immediately east of Balmoral (Gibson & Nihill, 1992). The western extent of the Glenelg Zone is obscured by Cenozoic cover immediately to the west of the Glenelg River at Dergholm. The portion of the Delamerian Fold Belt in Victoria is separated from that exposed in South Australia by a wide expanse of younger Murray Basin cover, which also covers the northern extent of the igneous–metamorphic complex. Migmatites and biotite–muscovite schists were drilled under the Murray Basin north of the exposed Glenelg River Complex rocks in VIMP-7, -12 and -13 (Maher *et al.,* 1997a). The Otway Basin covers the southern extent.

The timing of sedimentation is unknown, but presumed to be Cambrian or even Late Neoproterozoic. Protoliths of the Glenelg River Complex rocks, and less deformed and metamorphosed rocks further west, are correlated with the Moralana Supergroup (Preiss, 1982), which includes the Normanville and Kanmantoo Group metasediments in eastern South Australia. Many of the stratigraphic elements of the well-described Cambrian Normanville and Kanmantoo Groups in South Australia are present in parts of the Glenelg Zone, but mapping has yet to resolve their distribution. Stratigraphic continuity can only be established very locally due to the discontinuous outcrop and complex deformation. Therefore, the metasedimentary rocks have not been differentiated beyond Moralana Supergroup (Fig. 3.3), which encompasses the Normanville and Kanmantoo groups (Preiss, 1982). However, fault slices of tholeiitic to picritic basalt and gabbro that are extensive under cover north of the Glenelg Zone and have continuity into South Australia may be correlated with the Mount Arrowsmith Volcanics in westernmost New South Wales (Crawford *et al.*, 1997), the Truro Volcanics of the Normanville Group in South Australia, and the picrites of King Island, Togari Group metabasalts and Crimson Creek Formation rift tholeiites of northwestern Tasmania (Crawford & Berry, 1992). The correlation is on the basis of their linked distribution, their broadly similar age and the geochemical

Fig. 3.2: Geological domains and major faults of the Delamerian Fold Belt in western Victoria, showing the Glenelg Zone and Grampians–Stavely Zone, and location of the Glenelg River Complex west of the Yarramyljup Fault (modified from VandenBerg *et al.,* 2000)

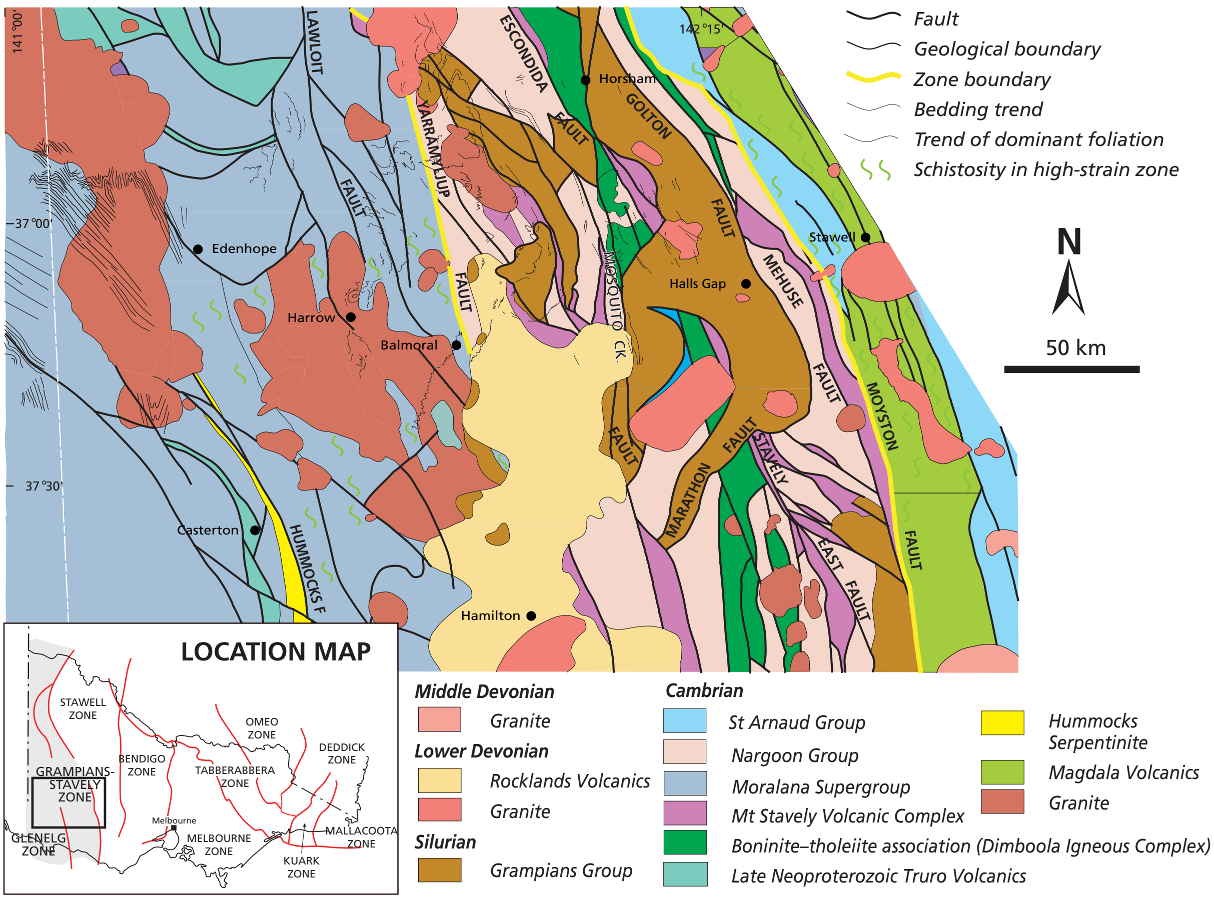


Fig. 3.3: Regional geological map of the Grampians area showing the distribution of the Mount Stavely Volcanic Complex rocks east of the Grampians (Mount Dryden Belt), south of the Grampians (Mount Stavely Volcanic Complex) and between the Grampians and the Yarramyljup Fault (Black Range and Glen Isla belts) (modified from VandenBerg *et al.*, 2000).

characteristics of basalts erupted in an evolving rift–drift setting (Crawford & Direen, 1998; Direen, 1999). Furthermore, several phlogopite-bearing cumulate ultramafic rocks in the western part of the Glenelg Zone are probably related to the same magmatic episode.

A structural history for the Glenelg Zone is resolved into five deformational events (see section 2.2.3). The second was the most intense and was responsible for the development of the most pervasive regional foliation (S_2), mesoscopic isoclinal folds and transposed layering.

Glenelg River Complex

The Glenelg River Complex (Fig. 3.4) has a NW–SE regional strike and is subdivided into (a) a southwestern metamorphic zone, (b) a northeastern metamorphic zone and (c) an axial granitic batholith zone. K–Ar mica and U–Pb zircon magmatic ages for several of the syn- to post-tectonic granites, and K–Ar metamorphic ages for several high-grade gneisses, all yield ages of about 520–500 Ma (Richards & Singleton, 1981; Turner *et al.,* 1993; Maher *et al.,* 1997a). These demonstrate that orogenic activity was part of the Cambro-Ordovician Delamerian Orogeny. Post-tectonic plutons give slightly younger K–Ar ages of about 500–480 Ma (Richards & Singleton, 1981; Turner *et al.,* 1993) and place an Early Ordovician age on the close of regional deformation. The long-held view of a link with the Delamerian Fold Belt in the Mount Lofty Ranges of South Australia (e.g. Wells, 1956) is confirmed by a similar deformation chronology and numerous geological similarities, such as comparable lithologies and distinctive post-tectonic granitic rocks. In particular, the metamorphic core of the Glenelg Zone has a similar structural and intrusive history to the strongly deformed and metamorphosed core of the southern Delamerian Fold Belt in the Mount Lofty Ranges (Sandiford *et al.,* 1992; Anderson & Gray, 1994; Foden *et al.,* 1999; Gray *et al.,* 2002).

Southwestern metamorphic zone

The southwestern metamorphic zone outcrops for about 40 km between the Wando Vale – Coleraine and Dergholm – Burke Bridge areas, prograding for about 15 km southwest to northeast through biotite, garnet–andalusite, sillimanite and migmatite zones (Fig. 3.4). VandenBerg *et al.* (2000) placed the western boundary of the Glenelg River Complex at the Hummocks Fault. Sweeping regional curvature of mainly NW-striking zone boundaries reflects F_5 folding. Low-grade rocks to the west of this fault are mainly turbiditic metagreywackes that retain sedimentary textures and structures. A sedimentary carbonate component is most apparent in minor outcrops of grey marble in Nolan Creek, and dark dolomitic slate occurs in Steep Bank Rivulet and the Glenelg River south of Dergholm (Wells, 1956). Rare dolomitic breccias are also present. Calc-silicates and actinolite–quartz schists in higher-grade zones appear to be derived from these dolomitic rocks.

In the low-grade part of the sequence there are diverse layer-parallel metabasites that were originally dolerite to gabbro and diorite (Wells, 1956; Gray *et al.,* 2002). These are now mainly plagioclase+actinolite+biotite and have textures that range from intact igneous, to severely deformed and finely layered, with foliated biotite or actinolite around plagioclase augen. Also within the low-grade rocks is a minor volcanic component. Rare metabasalts outcrop in the Dergholm–Nangeela area. They are fine-grained and plagioclase-phyric with weakly variolitic textures, and appear to be lava flows. Minor flows of plagioclase-phyric meta-andesite occur at Nolan Creek.

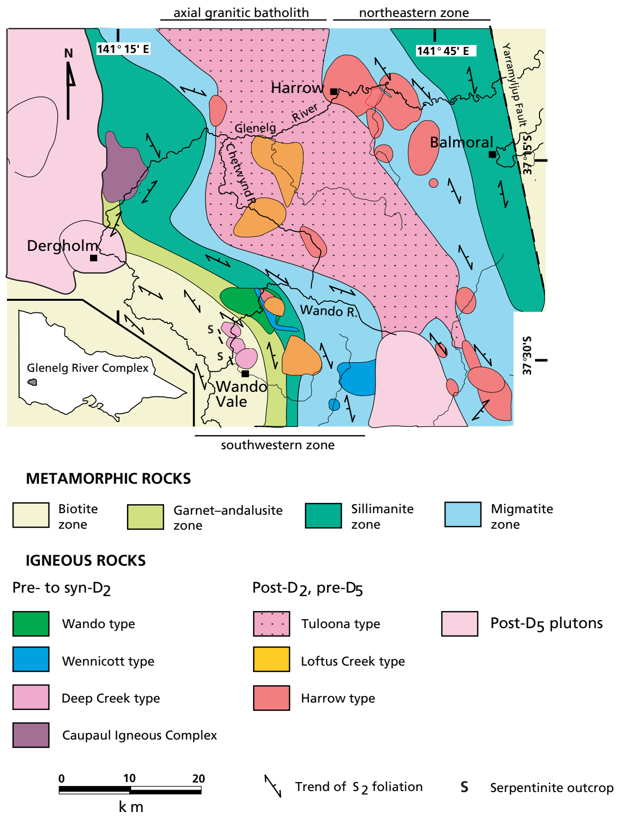


Fig. 3.4: Simplified Cambro-Ordovician geology of the Glenelg River Complex (adapted from Kemp et al., 2000). Depicted pluton shapes of the muscovite-bearing Harrow-type granitic rocks in the northeast are approximate only. Note that much of the area is covered by Permian and younger sedimentary and igneous rocks.

Fig. 3.5: Garnet–andalusite-zone schists with quartz-rich bands exposed in Corea Creek Gorge, near Wando Vale, southwestern metamorphic zone, Glenelg River Complex. Photograph by A. VandenBerg.

At biotite and higher grades, metabasites occur as two main types: fine-grained laminated amphibolites and coarse-grained metagabbros. Amphibolites are E-MORB compositions typical of tholeitic magmatism in extending continental crust (Gibson & Nihill, 1992; Anderson & Gray, 1994). Local proximity to, and textural gradation with, metagabbro suggest formation by recrystallisation of early gabbroic sills. Unrelated metagabbros with distinctively different MORB-type compositions (higher Al, Mg and Sr contents) form layer-parallel sheets up to 15 m thick at Wando Vale, in Steep Bank Rivulet and its tributaries (Gibson & Nihill, 1992; Turner *et al.,* 1993; Anderson & Gray, 1994) and along the Glenelg River between Dergholm and Burkes Bridge. It is not known if these are dykes or sills. Their actinolite+plagioclase metamorphic assemblage indicates that they may represent a distinct intrusive phase emplaced late in D_2.

Layer-parallel lenses of serpentinite occur as three large masses such as the Hummocks Serpentinite (100 x 350 m) (Wells, 1956; Turner *et al.,* 1993; VandenBerg *et al.,* 2000) and several 1–20-m thick sheet-like bodies. Their interiors preserve the textures of cumulate peridotites, and their margins often have a mylonitic fabric parallel to S_2 in the host rock. Relict chromites in the serpentinite have Cr/(Cr+Al) values (0.71–0.93) (Turner *et al.,* 1993), characteristic of highly depleted boninitic magmas rather than mid-ocean ridge basalt-type magmas, and are compositionally akin to chromites in Cambrian boninitic cumulates from further east in Victoria, such as at Heathcote and Howqua.

In areas showing medium- to high-grade metamorphism, the dominant rock type of the complex is homogeneous, grey, fine-grained quartzo-feldspathic schist with layers 10–100 cm thick (Fig. 3.5). Quartzo-feldspathic migmatites outcrop in places (Kemp & Gray, 1999b; Gray *et al.,* 2002). Calc-silicate rocks are similar throughout this grade range and vary from centimetre thickness to substantial units at least 150 m thick. Rocks are fine- to medium-grained, green to grey, with 10–60 cm internal layering; quartz-rich layers (<1 mm to >20 cm thick) are aligned with S_2.

There are numerous pre- or syn-D_2 intrusive rocks in the southwestern zone (Kemp *et al.*, 2002). A chain of plutons extending from the Wando River to Wennicott Creek includes, from west to east, the Wando Tonalite and Torah, Meissen and Deep Creek granodiorites (Anderson & Gray, 1994; Kemp *et al.*, 2002) and the Wennicott and Warradale tonalites (Bushell, 1996). Where present, igneous contacts and/or magnetic signatures indicate pluton diameters of 2–5 km. The S_2 fabric varies in intensity, with the Torah Granodiorite being gneissic, the Wennicott Tonalite having a pervasive biotite foliation, and the Deep Creek Granodiorite having a massive core and marginal foliation. These are subdivided into lithological types (Kemp *et al.*, 2002; Fig. 3.4). Wando-type intrusives are pale grey, foliated to gneissic hornblende tonalites with an even, medium grainsize that contain numerous igneous enclaves. Deep Creek types have poikilitic K-feldspar, sporadic hornblende and a distinctive high-Na, high-Sr character. Wennicott types commonly contain hornblende but, unlike Wando types, evolve towards lower K_2O with increasing silica (Kemp, 2002). In the Glenelg River valley about 10 km northeast of Dergholm is the small Caupaul Igneous Complex (Ferguson, 1993), composed of quartz diorite, diorite and gabbro-pyroxenite. The S_2 foliation intensity varies from weak to strong. Of a number of outcrops of gabbronorite and pyroxenite, three exceed 500 m across. Ferromagnesian minerals include orthopyroxene, clinopyroxene and anhedral hornblende, the latter commonly enclosing pyroxene. The range in textures, variable plagioclase abundance (0–20 %) and mafic nature of many rocks are consistent with a fractionating basaltic system. However, the limited variation in plagioclase composition indicates only part of the fractionation sequence is exposed. Deformational features are minor and timing of emplacement is unclear.

Several unfoliated medium- to coarse-grained Loftus Creek-type hornblende granodiorites with rare mafic igneous enclaves (Loftus Creek, Cloven Hills) also intrude the southwestern metamorphic zone.

Northeastern metamorphic zone

The northeastern metamorphic zone occupies a NW-striking belt about 15 km wide prograding from the sillimanite zone to migmatite grade from northeast to southwest. It extends westward from the Yarramyljup Fault, with the most complete section in the Glenelg River between Kanagulk and Harrow. The sillimanite zone is a monotonous sequence of quartzo-feldspathic and semi-pelitic schist with minor bands of biotite schist. The only prominent metapelite forms a 120-m wide band of quartz+plagioclase+biotite+muscovite+garnet+sillimanite schist in the Glenelg River near Kanagulk. Pegmatites are common in the sillimanite and migmatite zones as thin layers and irregular bodies. Metasedimentary rocks at migmatite grade are intermittently exposed along the Glenelg River, where they are intimately associated with varied granitic rocks (Kemp & Gray, 1999b; Kemp *et al.,* 2002). The proportion of leucosome increases westward leading to nebulitic migmatites, which grade rapidly into structurally concordant, muscovite-bearing plutons whose leucosome compositions correlate with those of adjacent migmatites. Comparable high-grade rocks occur in Bryan and Robson creeks about 30 km to the south along strike.

The northeastern metamorphic zone contains numerous granitic bodies, both structurally concordant and discordant. Textures are igneous, and structures in enclaves constrain emplacement to syn- or post-D_4 to pre-D_5. Structurally concordant plutons have gradational contacts with enclosing migmatites in which quartzo-feldspathic schist, migmatite, pegmatite and muscovite-bearing granitic rock are complexly interleaved at outcrop scale. In the Glenelg River about 1 km west of Scabbing Station Creek, nebulitic migmatite grades over about 50 m into Dunmore Leucotonalite. This body is 1.5 km across and is laden with migmatite enclaves and micaceous selvages, particularly at the margins; diffuse mafic schlieren define an internal fabric. In Schofield Creek, the Carrigeen Granodiorite is 750 m across with a core of pale, homogeneous, medium-grained, equigranular, muscovite granodiorite. Towards the margins it becomes progressively more heterogeneous, littered with micaceous selvages, microcline megacrysts and quartzo-feldspathic schist slabs. Locally it has a schlieric fabric defined by biotite+muscovite+ sillimanite. Ultimately, it grades into nebulitic migmatite at its northern and southern contacts (Fig. 3.6). Homogeneous, muscovite-bearing granitic rocks are also intrusive into the northeastern zone. The Harrow Granodiorite, exposed for about 3.3 km in the Glenelg River valley immediately east of Harrow, is medium-grained with primary muscovite, biotite, poikilitic microcline and sillimanite. It crosscuts adjacent units such as the Carrigeen Granodiorite and has a relatively low density of metasedimentary inclusions compared to the concordant bodies and a weak annular mica foliation. The Marn Mering Granodiorite, in the Glenelg River east of Schofield Creek, is 3 km across. It is a light grey, medium- to coarse-grained, porphyritic rock with microcline phenocrysts and generally lacks metasedimentary enclaves. Garnet–muscovite granitic bodies and felsic dykes are also widespread.

Axial granite batholith zone

The central parts of the Glenelg River Complex (Fig. 3.4) consist of a wide expanse of granites between Harrow and Chetwynd. These extend over 20 km across regional strike and separate the northeastern and southwestern metamorphic zones. Extrapolation from this transect using additional limited outcrop and aeromagnetic data implies a batholith at least 40 km long. This area is dominated by the numerous Tuloona-type granitic rocks (Tuloona, Chetwynd, Coojar and Patawilya, Glendara), which are unfoliated, grey granodiorite-granite with mafic igneous enclaves, minor muscovite and poikilitic microcline. Many of the plutons have marginal, muscovite-rich felsic phases that merge into migmatite (Fig. 3.7). Elongate quartzo-feldspathic schist enclaves define a flow structure. In addition, ellipsoidal, mafic, igneous-textured enclaves have plagioclase phenocrysts in a matrix of plagioclase, quartz and biotite.

Texturally distinctive, Loftus Creek-type hornblende granodiorites are medium- to coarse-grained with large (5 mm) euhedral biotites, poikilitic K-rich feldspar, prominent titanite euhedra and rare mafic igneous enclaves. There are four main plutons, all of which are unfoliated and post-date D_2. The Koolomurt Granodiorite crosscuts the Glendara Adamellite in Pigeon Ponds Creek. Further south the Cairns Creek Granodiorite is zoned from a porphyritic, hornblende-bearing margin to a felsic core without hornblende. The Loftus Creek and Cloven Hills granodiorites (Anderson & Gray, 1994; Kemp, 2002) intrude the southwestern metamorphic zone.

Undeformed post-tectonic granite occurs in the vicinity of Dergholm, just to the west of the metamorphic complex. Distinctive common features are red to buff colour, equidimensional grey to black quartz, highly perthitic microcline, albitic plagioclase, common accessory fluorite, pleochroic biotite (very dark brown to black), and evolved chemical compositions. Turner *et al.* (1993) treated these rocks as a single body, the 'Dergholm Granite'. However, three or four textural types are recognised and regarded as forming distinct plutons, with their shapes deduced from magnetic signatures. Magnetic data also indicate continuity as a batholith extending subsurface into South Australia, confirming petrological links to scattered exposures as far distant as Murray Bridge. The Baileys Rocks Granite outcrops over 8 km in a NNW direction in the vicinity of the reserve of the same name. It has a lobe-like magnetic shape about 7 km across extending N–S for at least 15 km. It is characterised by buff K-rich feldspar phenocrysts and may contain hornblende and titanite. The Dergholm Granite, equidimensional and 7 km across, is located in the environs of the Glenelg River around Dergholm. It has a distinctive medium-grained, equigranular texture with beta-quartz crystals set in square, buff crystals of both K-rich and plagioclase feldspar. An oval pluton with low magnetic intensity and diameter of 12 km at Poolaijelo is defined as the Poolaijelo Granite. Minor exposures in and around Salt Creek, south of Poolaijelo, are distinct from the Baileys Rocks lithology. The most abundant type is even-textured and medium- to coarse-grained, having equidimensional quartz grains combined with feldspar, usually uniformly red, but sometimes contrasting with cream or pale green. Local variants are finer grained and grade to aplite.

Age constraints and correlations

Most of the Glenelg Zone is probably correlated with the Late Neoproterozoic – Early Cambrian Moralana Supergroup and its correlatives in the Koonenberry belt of westernmost New South Wales. These sequences are dominated by turbiditic metasediments and shales, with minor carbonates,

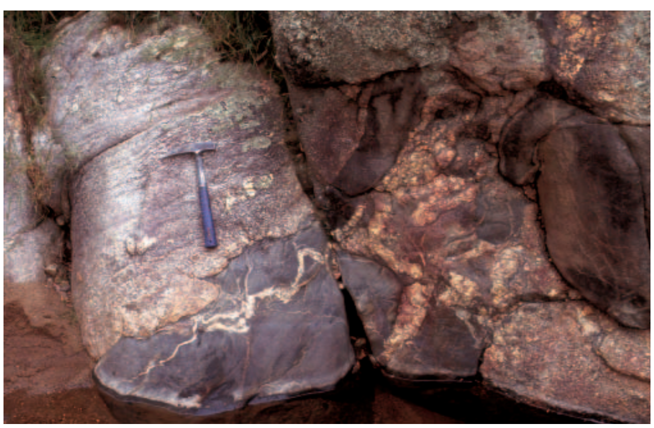


Fig. 3.6: Transition zone from schist and migmatite to muscovite granite exposed in Schofield Creek, near Harrow, northeastern metamorphic zone, Glenelg River Complex. Photograph by R. Cayley.

Fig. 3.7: Marginal phase of the Tuloona Granodiorite exposed in Schofield Creek, with a directional fabric defined by elongate metasedimentary enclaves and micaceous schlieren. An igneous-textured microgranular enclave occurs in the centre of the photograph near the pen. Photograph by T. Kemp.

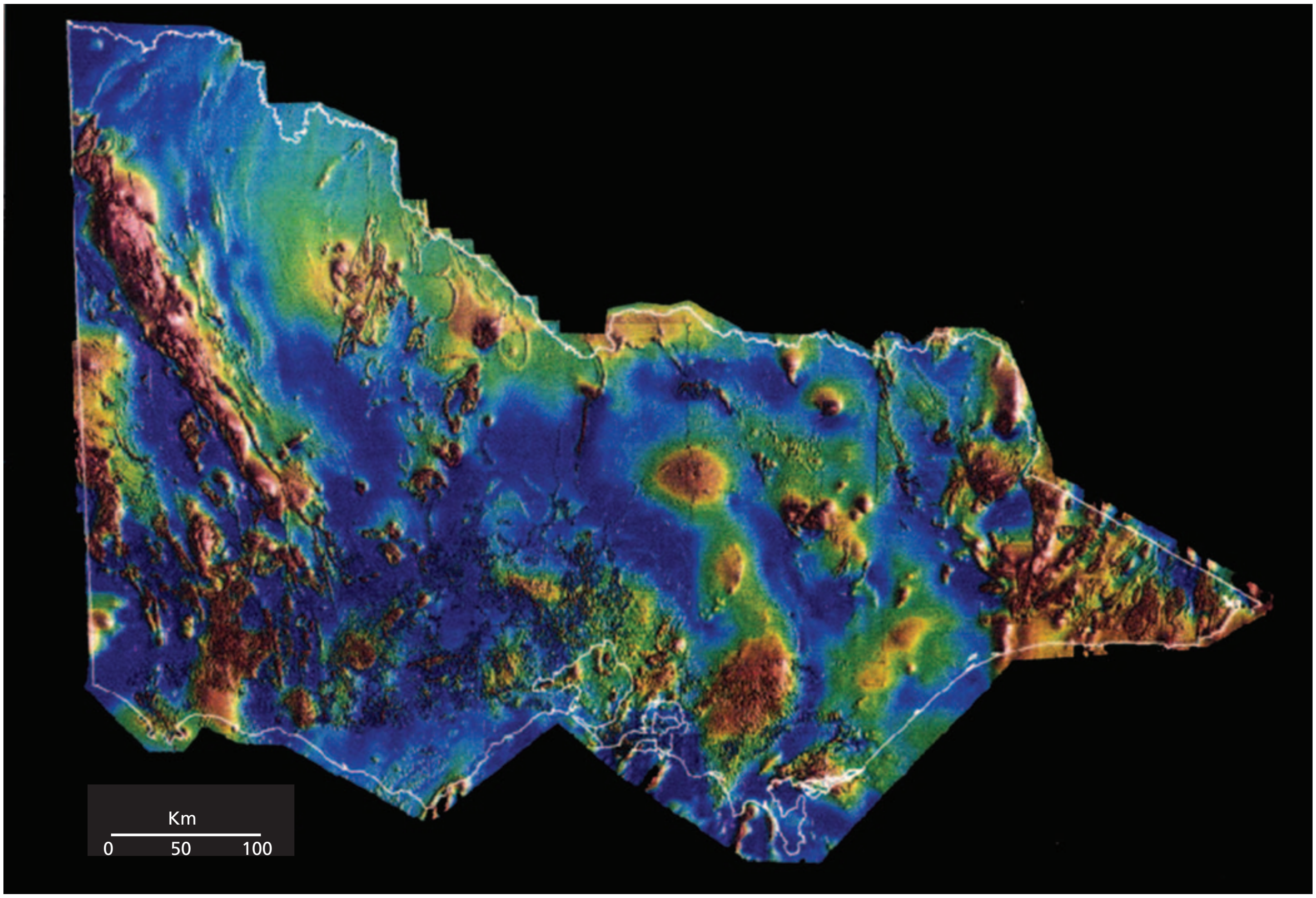


Fig. 3.8: Aeromagnetic image of Victoria showing the major magnetic high striking northwest beneath the Murray Basin. Known as the Dimboola Igneous complex of VandenBerg *et al*. (2000), the rocks responsible for this feature may be either Late Neoproterozoic metabasalts of the 600-Ma volcanic passive-margin sequence (Crawford & Direen, 2001) or an oceanic forearc-derived basaltic pile that collided with the passive margin in the early Middle Cambrian. (image courtesy of Geological Survey of Victoria).

with a major pulse of rift-type transitional alkaline to tholeitic magmatism at about 600–590 Ma (Crawford *et al.,* 1997). Due to a lack of fossils, age constraints on the rocks of the Glenelg Zone consist of radiometric age determinations on detrital and metamorphic minerals, and also on the numerous Late Cambrian granites which intrude the sequence. A relatively discrete population of detrital zircons at about 590 Ma in biotite gneiss from drillhole VIMP-12 (Maher *et al.,* 1997a) provides a maximum possible age for the precursors of the Moralana Supergroup rocks here. A minimum age constraint is provided by the Early Cambrian Bringalbert Gabbro in the same drillhole (524±9 Ma; Maher *et al.,* 1997a). This appears to have intruded the sedimentary sequence, and suffered some subsequent deformation. Thus, the age of at least part of the pelitic sequence in this region is broadly bracketed at between 590 and 524 Ma, i.e. Ediacaran to Early Cambrian.

K–Ar cooling ages of granitic rocks in the Glenelg River Complex are late Middle Cambrian (~500 Ma) for syntectonic pegmatite and Late Cambrian (~490 Ma) for post-tectonic pegmatite and the post-tectonic A-type Baileys Rocks Granite (Turner *et al.,* 1993). Cooling of the metamorphic rocks to about 300ºC (blocking temperature of K–Ar biotite) also occurred at this time (Turner *et al.,* 1993; Maher, *et al.,* 1997). Two Ar/Ar dates on metamorphic biotite are also Late Cambrian (Turner *et al.,* 1993). Thus the sedimentary protoliths are older than Late Cambrian, when the deformation and accompanying metamorphism appear to have occurred.

3.2.3 Grampians–Stavely Zone

The Grampians–Stavely Zone (Fig. 3.2) is distinguished from the Glenelg Zone by lower grade, less-deformed rocks, and by the absence of syn-tectonic granites. The Cambrian rocks of this zone are generally poorly exposed, largely buried beneath the spectacular younger cover of the Grampians Group. They consist of a number of belts of calc-alkaline volcanics (the Mount Stavely Volcanic Complex, made up of the Black Range and Mount Dryden belts and correlated units) that are structurally intercalated with black shale and sandstone of the Nargoon Group (Fig. 3.3). In the western half of the zone, these rocks are of greenschist grade and possess a cleavage, but east of the Escondida Fault they are of lower grade and lack cleavage development. There are scattered occurrences of tholeiitic to boninitic lavas and intrusives throughout the zone, with most occurring as narrow fault slices discovered by drilling beneath shallow cover. The Williamsons Road Serpentinite occurs as elongate faulted slivers of boninitic cumulate-derived serpentinite within calc-alkaline lavas of the Mount Stavely Volcanic Complex. Aeromagnetic data, however, show a major belt of magnetic rocks (Fig. 3.8) — the Dimboola Igneous Complex (VandenBerg *et al.*, 2000) — becoming extensive to the north under Murray Basin cover. Whether this represents massive accumulations of the tholeiite-boninite sequence (the Dimboola Igneous Complex of VandenBerg *et al.,* 2000), or seaward-dipping reflector packages forming part of the 600-Ma rifted margin of southeastern Australia (Crawford & Direen, 1998; Direen, 1999; Crawford & Direen, 2001) remains to be established.

Mount Stavely Volcanic Complex

The Mount Stavely Volcanic Complex (Buckland & Ramsay, 1982; Buckland, 1986; Crawford, 1982, 1988; Donaghy, 1994; Crawford *et al.,* 1996a, b; Cayley & Taylor, 1997) forms a series of NW-trending linear volcanic belts in the Grampians–Stavely Zone (Fig. 3.3). These are the subparallel Mount Stavely and Mount Dryden belts, respectively south and east of the Grampians. The two belts outcrop sporadically as a number of low hills with good exposure. Just west of the Grampians in the Black Range, several belts of volcanics are much more poorly exposed but have been extensively drilled. These lavas have the same calc-alkaline geochemistry as the other belts. Although individual belts show considerable petrographic and geochemical variation along strike, the variation between belts strongly overlaps (Buckland, 1986; Donaghy, 1994; Direen, 1999). The belts of volcanics incorporate rare fault slices of variably serpentinised boninitic rocks and are intercalated with low-grade and weakly deformed turbidites and black shale of the Nargoon Group. Most of the volcanic rocks are essentially undeformed and only partly altered by prehnite-pumpellyite grade metamorphism, so that original igneous textures are largely preserved.

The Mount Stavely Volcanic Complex also outcrops south of the Grampians, where it consists of fault-bounded but internally undeformed blocks of volcanics. The basal sequence is dominated by medium- to high-K, calc-alkaline andesite and dacite of the Fairview Andesitic Breccia (Table 3.1). More felsic compositions in the upper parts of the pile include the Narrapumelap Road Dacite and the Nanapundah and Towanway Tuffs. Small tonalite-trondhjemite plutons of the Lalkaldarno Tonalite (Fig. 3.9c) have intruded the volcanic rocks. Detrital zircons in a volcaniclastic rock, and magmatic zircons in a meta-dacite of the Mount Stavely Volcanic Complex, have yielded crystallisation ages of 501±9 Ma and 495±5 Ma respectively (Stuart-Smith & Black, 1994). Biotite from the Lalkaldarno Tonalite yielded an Ar/Ar age of 500±2 Ma (Foster *et al.,* 1996b).

Mount Dryden Belt

The Mount Dryden Belt outcrops intermittently between the Moyston and Mehuse Faults, east of the Grampians (Fig. 3.3). Principal sites where volcanics in the belt outcrop are (from north to south) Mount Dryden, Lake Fyans, McMurtrie Hill, Jallukar, Barton and much further south at Lake Bolac. Mount Elliot occurs halfway between the Mount Dryden Belt and the Mount Stavely Belt as an isolated hill on a NW-trending linking belt. Parts of the Mount Dryden Belt have previously been described in some detail by Buckland (1986) and Crawford (1988). The width of the Mount Dryden Belt is greater than previously thought, up to 5 km at Mount Dryden. Magnetic data show that the isolated outcrops expose different levels of a conformable west-facing sequence within the belt, rather than being direct along-strike extensions. The belt has been informally subdivided into three major units that are intercalated within it (Cayley & Taylor, 2001). These are:

1. About 1000 m of relatively low-Ti andesitic to dacitic lavas including pillow lavas (Table 3.1) and high-level intrusions;
2. About 500 m of volcanic conglomerate (Fig. 3.10);
3. More than 500 m of volcaniclastic sandstone.

It is impossible to determine the original thicknesses of the units because of faulting; however, the minimum true thickness of the conformable sequence at Mount Dryden is 2000 m, with a possible further 1500 m of undifferentiated volcanics to the west, beneath thin alluvial cover. In the vicinity of McMurtrie Hill, these volcanic and volcaniclastic rocks have been intruded by a diorite sill with a thickness of 450 m, but it appears to be truncated on its eastern side by the faulted eastern margin of the belt.

Black Range Belt

In the Black Range about 35 km west of Halls Gap, midway between the Grampians and the Yarramyljup Fault, calc-alkaline volcanics outcrop in the 25-km long, NW-trending Black Range Belt (Fig. 3.1). A smaller (<10 km strike length) belt, the Glenisla Belt, occurs to the east of the main belt. A third belt, the Tyar Belt, occurs beneath laterite to the west of the Black Range and has been delineated by aeromagnetics and drilling (Spencer-Jones, 1965; McArthur, 1990; Cayley & Taylor, 1997). Mafic to intermediate volcanics occur in all three belts and intermediate to felsic volcanics occur on the eastern side of

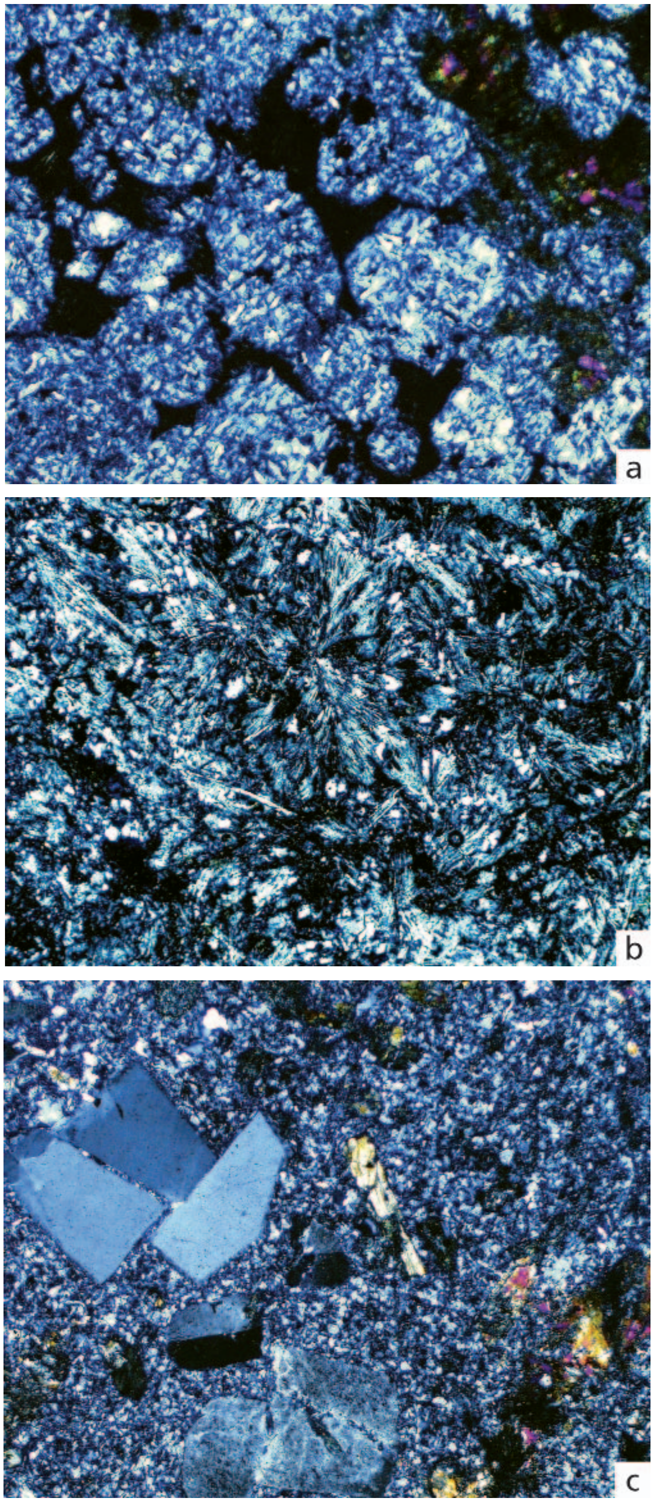


Fig. 3.9: Photomicrographs of rocks from the Mt Stavely Volcanic Complex. (a) Strongly altered base of an olivine-rich picritic lava flow showing serpentinised euhedral former olivine crystals, some with small chromite inclusions, and glassy groundmass now replaced by low birefringent chlorite. From Yanac South, Western Victoria, prospect drilled by MIM. (b) Quenched top of a dacitic lava with typical rosettes of rapidly cooled plagioclase, and an altered and recrystallised, formerly glassy groundmass now replaced by chlorite, quartz and albite. From McRaes prospect, Black Range, Western Victoria. (c) Shallow intrusive tonalite of the Lalkaldarno Porphyry from south of Mt Stavely, showing euhedral hornblende, quartz and albitised plagioclase phenocrysts. All images are c.5 mm across and are taken with crossed polars.

Fig. 3.10: Volcanic conglomerate/breccia, Mount Dryden Volcanics, north slope of Mount Dryden. Photograph by A. VandenBerg.

the main belt (Cayley & Taylor, 1997). Unlike the other belts further east, the volcanics in the Black Range show variably developed cleavage or schistosity.

Drill holes bored by CRAE at McRaes prospect and elsewhere in the Black Range Belt intersected diverse volcanics dominated by dacitic lavas (Fig. 3.9b), but including quartz-phyric rhyolite and subordinate andesitic and basaltic lava (Direen, 1999). Similar rocks have been drilled at VIMP-3 during the Geological Survey VIMP drilling program in this region. Analytical data for andesitic lavas from these drill holes were reported by Maher *et al.,* (1997a) and Direen (1999) (see Table 3.1). The plagioclase+augite-phyric andesites and dacites of the Black Range Belt have medium-K calc-alkaline affinities, with characteristically low TiO_2 values and other compositional features indicating a possible correlation with the Mount Dryden Belt (Maher *et al.,* 1997a; Direen, 1999).

Setting and regional correlation of the Mount Stavely Volcanic Complex.

The Mount Stavely Volcanic Complex shares similar submarine eruptive setting, geochemistry and age to the Mount Read Volcanics of Tasmania (Crawford *et al.,* 1996a). In Tasmania, these rocks are interpreted as post-collisional rift

	1	2	3	4	5	6	7	8	9	10	11
	298.8 m	**216.6 m**	**FP1**	**FP4**	**GM47-227**	**GM47-70.8**	**GM48-96.2**	**26226**	**26222**	**26203**	**26205**
SiO_2	44.70	46.40	52.50	54.80	62.70	75.00	72.60	56.40	59.60	58.60	61.40
TiO_2	0.05	0.16	0.31	0.26	0.86	0.34	0.60	0.33	0.31	0.50	0.49
Al_2O_3	3.91	11.30	13.10	11.30	15.40	12.40	13.40	15.30	14.60	15.30	15.60
FeO*	10.40	9.54	9.96	7.89	6.37	2.63	5.23	10.20	7.95	8.91	9.42
MnO	0.17	0.22	0.16	0.13	0.13	0.04	0.17	0.10	0.11	0.15	0.12
MgO	39.30	22.60	14.30	14.80	4.49	3.21	3.71	7.70	5.34	3.91	2.91
CaO	0.24	7.73	5.66	5.90	3.12	0.32	0.33	4.03	8.76	7.58	5.16
Na_2O	0.06	0.89	1.13	0.81	2.12	4.60	1.91	5.52	2.36	4.61	4.03
K_2O	0.01	0.02	1.72	3.19	3.99	1.03	1.29	0.37	0.83	0.39	0.70
P_2O_5	0.00	0.01	0.07	0.05	0.11	0.07	0.17	0.03	0.08	0.12	0.13
LOI	11.90	6.76	10.10	7.10	4.81	2.18	3.71	1.76	1.43	1.78	1.01
Trace elements (ppm)											
Ni	1567	976	429	376	26	15	8	51	45	35	17
Cr								229	186	59	18
V	68	164	199	178	157	64	90	178	197	222	263
Sc	17	34	36	33	25	8	13	34	27	25	22
Zr	1	7	28	25	131	116	114	48	56	60	69
Nb	<1	<1	1.3	1.7	4	2.3	2.4	<2	<2	<2	<2
Y	1.3	8.3	8	8	23	14	12.5	8	10	12	13
Sr	3	21	122	111	81	66	67	74	246	235	258
Rb	<1	<1	20	44	64	17	9	15	22	14	18
Ba	4	5	648	859	522	236	382	26	96	121	29
	12	**13**	**14**	**15**	**16**	**17**	**18**	**19**	**20**	**21**	**22**
	26229	**108.7 m**	**V2S1**	**V2O2**	**V2A**	**V2K**	**V2Q**	**V2W2(2)**	**V2I3**	**V2H3**	**26233**
SiO_2	69.40	63.60	53.30	58.80	59.80	60.80	63.00	65.40	72.20	75.00	65.80
TiO_2	0.43	0.43	0.75	0.73	0.61	0.81	0.97	0.58	0.48	0.46	0.54
Al_2O_3	13.90	14.50	15.40	15.60	12.50	16.90	16.20	12.50	13.80	13.50	16.80
FeO*	5.30	6.87	10.90	7.67	8.53	5.76	5.48	6.80	2.99	2.59	2.93
MnO	0.14	0.09	0.16	0.12	0.14	0.09	0.08	0.09	0.05	0.04	0.05
MgO	1.54	3.06	6.03	5.04	7.07	3.55	2.79	4.27	1.21	0.46	3.73
CaO	3.22	4.32	10.05	6.94	7.29	6.36	6.35	5.89	2.89	0.35	4.73
Na_2O	4.37	3.25	2.39	3.81	2.87	4.86	4.23	3.47	3.65	4.99	4.89
K_2O	1.52	0.49	0.75	0.98	0.96	0.62	0.63	0.73	2.63	2.59	0.49
P_2O_5	0.19	0.19	0.29	0.28	0.24	0.26	0.31	0.22	0.12	0.07	0.13
LOI	1.03	3.15	6.20	5.16	7.22	3.64	2.87	4.36	1.26	0.50	3.78
Trace elements (ppm)											
Ni	4	12									88
Cr	15		123	99	313	59	4	52	25	3	59
V	188	185	349	220	213	161	173	159	87	28	83
Sc	18	22	36	31	35	22	19	23	12	11	10
Zr	85	55	43	137	97	168	109	125	138	171	87
Nb	<2	2	2	5	4	4	4	3	7	4	
Y	22	12	18	14	12	15	16	12	20	26	15
Sr	416	50	430	418	407	525	733	466	263	151	451
Rb	36	11	17	24	25	15	14	22	56	39	12
Ba	330	273	191	149	160	106	86	119	489	499	108

Table 3.1: Whole-rock analyses for Late Neoproterozoic and Cambrian greenstones from the Stavely Greenstone Belt and Glenelg Zone. 1, 2: Dunitic cumulates associated with Late Neoproterozoic picritic lavas, Yanac South drillholes YANS1 and YANS2 drilled by MIM; depths shown. 3, 4: Boninitic lavas from the early Middle Cambrian allochthon drilled at the Frying Pan prospect by CRA, near Jallukar, within the Stavely Greenstone Belt. 5–22 are all from the Mt Stavely Volcanic Complex. 5–7: Dacite and rhyolite lavas drilled by CRA at McRaes prospect in Black Range. 8–11: Low-Ti andesites from Mt Dryden. 12: Dacite from Jallukar. 13: Low-Ti andesite akin to those at Mt Dryden, drilled in VIMP-3 by Geological Survey of Victoria (Maher *et al.,* 1997a). 14–21 are from the Mt Stavely Volcanic Complex, analyses 12–19 are from andesites and dacites of the Fairview Andesitic Breccia, and 20 and 21 are from the Narrapumelap Road Dacite. 22: Tonalite from the Lalkaldarno Tonalite intruding the Mt Stavely Volcanic Complex. 1–13 and 22 are from A. J. Crawford (unpublished) and Direen (1999). 14–21 are from Donaghy (1994).

volcanics erupted into rift basins developed in the older crust during the waning phases of the Delamerian/Tyennan Orogeny (Crawford & Berry, 1992). The strong temporal and compositional similarities between the post-collisional volcanics in western Victoria and better exposed correlatives in western Tasmania, suggest there is some potential for VHMS- and porphyry-style mineralisation in the unexposed sequences that have been shown by magnetics to extend for some considerable distance beneath the Murray Basin.

Nargoon Group

This group incorporates all the sedimentary bedrock of the Grampians–Stavely Zone, comprising the poorly outcropping black slate and sandstone in the Black Range area as well as the much better exposed and described Glenthompson Sandstone further east (Buckland, 1986). The sedimentary rocks are all interpreted to have originally overlain the Mount Stavely Volcanic Complex, and thus represent the sedimentary fill in the upper parts of the rift basin occupied by those volcanic rocks.

The Glenthompson Sandstone, originally defined in the vicinity of the Mount Stavely Volcanic Complex (Buckland, 1986), forms a uniform and widespread package of micaceous quartz-rich turbidites which comprise much of the sedimentary bedrock in the Grampians–Stavely Zone between the Moyston and Escondida faults (Fig. 3.3). This includes the discontinuous belt of turbiditic sandstone referred to as the Moyston Sandstone by Watchorn & Wilson (1989), sandwiched between this fault and the Mount Dryden Belt.

The Glenthompson Sandstone possibly conformably overlies Mount Stavely Volcanic Complex. A gradational sedimentary contact with the underlying volcanic rocks is indicated by the presence of interbedded rhyolitic and andesitic volcaniclastic arenite beds within the Glenthompson Sandstone (e.g. Stuart-Smith & Black, 1994; Donaghy, 1994).

Between the Escondida and Yarramyljup faults, the Nargoon Group consists of extremely poorly exposed quartz-rich turbiditic metasandstone, mudstone, schist and black slate, which separate the fault-emplaced belts of the Mount Stavely Volcanic Complex (Cayley & Taylor, 1997). Apart from minor outcrops in the vicinity of the Black Range, south of Balmoral, and in the vicinity of Chatsworth, most information on these rocks comes from drill holes intersecting the sequence in the Black Range region. From the limited data available, it appears that they are predominantly deep-marine terrigenous to hemipelagic metasedimentary rocks of low greenschist grade, although they reach biotite grade in the vicinity of Chatsworth (Stuart-Smith & Black, 1994). Weak magnetic striping in the various metasedimentary rocks of the Nargoon Group may be due to the presence of dykes, or interbedded mafic volcanics, as interpreted to the west of Mount Stavely (Stuart-Smith & Black, 1994).

Although unfossiliferous, the age of the Nargoon Group is well constrained to Late Cambrian by the underlying 500 Ma volcanics of the Mount Stavely Volcanic Complex, and the 489 Ma Bushy Creek Granite which intrudes the deformed Nargoon Group just west of these volcanics. The tight timing of deposition, deformation and intrusion of the group suggests that it was deposited during the waning phases of the Delamerian Orogeny. This inference is confirmed by the large population of 500–510 Ma detrital zircons in the unit, which must have been derived from Delamerian magmatism as recorded in the Glenelg Zone to the west.

Mafic–ultramafic rocks

Scattered occurrences of ultramafic to mafic, boninitic and tholeiitic volcanics and intrusive rocks occur in the eastern portion of the Grampians–Stavely Zone. On the basis of their distinctive geochemical signature, they are unambiguously correlated with the main Cambrian tholeiite–boninite association of the Heathcote Greenstone Belt in the Lachlan Fold Belt further east, and also with the Early Cambrian mafic–ultramafic complexes of western Tasmania (Crawford & Berry, 1992). Ultramafic serpentinite slivers, such as the Williamsons Road Serpentinite, in-faulted into the Mount Stavely Volcanic Complex are included in this suite, as are unusual boninitic lavas at the Frying Pan prospect west of Moyston (Menpes, 1994), and at Wartook west of the Grampians (Stewart, 1993). The presence of these presumed fault slices west of the Moyston Fault-defined margin of the Lachlan Fold Belt is important. It implies that the tholeiite–boninite basement of the southeastern Lachlan Fold Belt was either originally thrust westwards onto Grampians–Stavely Zone elements of the Delamerian Fold Belt, or was thrust west of the Moyston Fault at some late stage of the Delamerian deformation event.

Widespread and voluminous magnetic mafic rocks beneath younger cover rocks north of the Grampians are indicated by aeromagnetics (Fig. 3.5) and limited drilling. The large magnetic package around Dimboola has been drilled by North Ltd. and shown to include altered basaltic lapilli tuff and basalt, with microgabbro and pyroxenite; geochemical data for these rocks are unavailable (O'Neill, 1994). Further south, the VIMP-9 drill hole intersected moderately augite-phyric tholeiitic basalt with hyaloclastite peperitic interaction with the interlayered red siltstone. (Maher *et al.,* 1997a). A smaller magnetic high close to the Victoria–South Australia border at Yanac South was drilled by MIM and intersected distinctive olivine-rich picritic lavas and cumulates with rift-tholeiite compositions (Direen, 1999) (Fig. 3.9a). These have been correlated with the Late Neoproterozoic picrites on King Island and in the Smithton Trough in northwestern Tasmania.

3.2.4 Regional synthesis

A geological synthesis of the Glenelg Zone is hampered by a lack of age control because the sediments are unfossiliferous and the mafic volcanics lack dateable minerals. Much of the regional synthesis draws on lithological correlations with better-understood sequences in South Australia and Tasmania that were also involved in the Delamerian Orogeny. In the Glenelg Zone, the mainly quartzo-feldspathic turbidites and associated mafic volcanics represent an outboard portion of the passive margin sequence of the Delamerian Fold Belt, referred to as the Stansbury Basin (Belperio *et al.,* 1988). Basalts in easternmost South Australia and westernmost New South Wales (Crawford *et al.,* 1997) are compositionally transitional from intraplate basalts to rift tholeiites, and record magmatism preceding and during breakup of this section of Gondwana, at about 600 Ma. East of the Glenelg Zone, the large belt of magnetic mafic igneous rocks may represent the remnants of an arc–forearc complex that lay to the east of the Stansbury Basin in the Cambrian (VandenBerg *et al.,* 2000). An alternative interpretation for these rocks is that they represent rift tholeiites formed originally as seaward-dipping reflectors on an east-facing volcanic passive margin during the 600–590 Ma continental breakup in eastern Gondwana (Crawford & Direen, 1998; Direen, 1999).

Serpentinised ultramafic cumulates and comagmatic lavas with boninitic affinity occur as fault-bounded slices referred to as the Dimboola Igneous Complex by VandenBerg *et al.* (2000). These appear to be associated in places with MORB-type tholeiitic lavas. This association is better exposed and well studied from the Heathcote and Mount Wellington Greenstone Belts of the Lachlan Fold Belt further east, where it has been interpreted as part of a massive allochthonous sheet of crust and upper mantle of Early Cambrian age that was emplaced westward onto the 600-Ma passive margin (Crawford & Berry, 1992). This arc–continent collision represents the earliest phase of the Delamerian Orogeny, and subsequent crustal thickening led to uplift of the Glenelg Zone, which shed sediment into one or more localised extensional submarine rift basin(s) forming during the uplift. These were initially filled with calc-alkaline volcanics of the Mount Stavely Volcanic Complex and related suites in the Mount Dryden and Black Range Belts. As volcanism waned the marine sandstones and black shale of the Nargoon Group accumulated upon the volcanics and were only weakly deformed shortly after deposition in the terminal phases of the orogeny. Based on (1) the age and compositional similarity of all but the Mount Dryden Belt lavas with the Mount Read Volcanics in Tasmania, which unambiguously post-date emplace-ment of the forearc-derived ophiolitic allochthons in western Tasmania and Victoria (Crawford & Berry, 1992), and (2), the occurrence of in-faulted slices of boninite-derived serpentinised ultramafics, it is assumed here that these calc-alkaline volcanic suites in western Victoria are also post-collisional magmatic suites. The Mount Dryden Belt remains undated, but regional geological considerations (e.g. it is conformably overlain by Late Cambrian Glenthompson Sandstone) suggest that these lavas too are a post-collisional suite.

3.3 Lachlan Fold Belt

3.3.1 Stawell and Bendigo zones

In the western Lachlan Fold Belt the Magdala, Pitfield and Heathcote Volcanics occur in the hanging wall of major fault zones (Fig. 3.1). These three predominantly volcanic packages are inferred to be exposed portions of a sheet of ocean-floor volcanics that regionally underlie the Cambro-Ordovician turbidite pile. The major craton-directed Moyston Fault has thrust the mid-crustal levels of the Lachlan Fold Belt over the Delamerian Fold Belt, and brought the Moornambool Metamorphic Complex to the surface (VandenBerg *et al.,* 2000). This metamorphic complex contains a variety of mafic and pelitic schists derived from Cambrian volcanic and sedimentary protoliths during the Late Ordovician – Early Silurian Benambran Orogeny (see Chapter 2, section 2.3.1). The unfossiliferous and presumed Cambrian turbidites of the St Arnaud Group overlie the volcanics in the west, whereas the Goldie Chert and Knowsley East Shale in the east lie between the Cambrian volcanics and the Ordovician turbidites of the Castlemaine Group.

Magdala Volcanics

The Magdala Volcanics occur in the Moornambool Metamorphic Complex, a 15-km wide high strain zone adjacent to the Moyston Fault (Cayley & Taylor, 1998b). The volcanics consist of lavas and volcaniclastic sediments, best known from the Stawell Mine ('Footwall volcanics' of Watchorn & Wilson, 1989 and 'Magdala volcanogenics' of Watchorn, 1986). Other exposures of this unit occur south of Great Western and southwest of Ararat. The Magdala Volcanics consist of massive and pillowed basaltic lavas, volcaniclastic interflow sediments and minor chert, attesting to eruption in a submarine environment. The volcanics are regarded as the protolith of the higher grade Deenicull Schist and Carrolls Amphibolite, which also outcrop within the Moornambool Metamorphic Complex, such as in the Moyston area.

The Magdala Volcanics form the lowest and oldest rock unit exposed in the westernmost portion of the Lachlan Fold Belt. They underlie the St Arnaud Group but the base is not exposed and their thickness is not known. The Magdala Antiform has a width of about 1000 m of deformed basalt in the Stawell gold mine, but there is evidence of considerable structural thickening here (Watchorn & Wilson, 1989). The thickest package that is relatively undeformed is approximately 200 m thick in a fault slice at Sheepfold Hill, southwest of Ararat. At Stawell, a 10–70 m thick sequence of layered and massive volcaniclastic sediments occurs on the west flank of the Magdala Antiform (Willman, 1987; Watchorn & Wilson, 1989; Phillips *et al*., 2002). An apparently conformable transition from lava through volcaniclastic sediments up into the overlying Warrak Formation of the St Arnaud Group is preserved in some drillholes on each side of the Magdala Antiform, although disrupted by some faults (Watchorn, 1986; Willman, 1987; Fredrickson & Gane, 1998).

Unpublished geochemical studies of the Magdala Volcanics metabasalts show them to be tholeiitic basalts essentially identical in composition to the better exposed Cambrian tholeiitic basalts in the Heathcote and Mount Wellington Greenstone Belts (Crawford, 1994). Whole-rock geochemical analyses (Table 3.2) show a strong Fe enrichment, and increases in TiO_2 and V with advancing fractionation, features typical of strongly differentiated tholeiitic suites (Will, 1990). The geochemical signature is transitional between ridge-generated basalts and island-arc tholeiites, similar to rocks erupted in modern backarc basin settings. Some samples from the Magdala Volcanics have extremely low TiO_2 contents and are depleted in Zr and Y (Rowe, 1989), features that are atypical of the tholeiitic rocks at Stawell, but that are very close to the low-Ti tholeiites in the Early Cambrian mafic–ultramafic complexes in western Tasmania (Brown & Jenner, 1989; Crawford & Berry, 1992). Some samples of strongly deformed Carrolls Amphibolite contain relict Cr-rich and Al-poor chromite grains, which together with the widespread occurrence of tremolite and magnesian hornblende in the Deenicull Schist and the Carrolls Amphibolite (Cayley & Taylor, 2001) suggest a substantial boninitic component to parts of the Magdala Volcanics sequence. This tholeiite–boninite association is identical to the better-exposed Heathcote Volcanics and Dookie and Thiele igneous complexes.

The Magdala Volcanics are economically important, because the lavas form the rigid buttress of rock at Stawell against which quartz-gold mineralisation has developed. The overlying volcaniclastic sediments host some of the most mineralised parts of the Central Lode System (Fredrickson & Gane, 1998). Strongly metamorphosed equivalents of the Magdala Volcanics also host gold mineralisation at Moyston. The Mount Ararat copper lode may represent a deformed and metamorphosed VHMS deposit related to the Magdala Volcanics, in the higher-grade parts of the Moornambool Metamorphic Complex south of Ararat.

Because the Magdala Volcanics underlie the St Arnaud Group, they are therefore no younger than Cambrian. An Early Cambrian age is inferred by comparison with similar rocks exposed at Heathcote, where there are associated Early Cambrian fossils. A Pb/Pb isochron age of 700±30 Ma

	1	2	3	4	5	6	7	8	9	10	11	12	13
	68890	**68887**	**68888**	**P3**	**P4**	**BH7**	**BH12**	**BH11**	**BH6**	**26255**	**26263**	**26265**	**26266**
SiO_2	49.60	51.60	51.30	48.66	49.17	46.70	49.80	49.80	51.50	62.50	64.00	65.00	65.00
TiO_2	1.25	1.56	2.11	1.07	0.98	0.03	0.06	0.12	0.12	0.30	0.26	0.28	0.25
Al_2O_3	16.60	14.80	13.80	16.72	15.50	18.00	16.80	16.70	16.10	13.00	15.10	16.30	16.80
FeO*	10.30	9.90	11.30	12.25	11.28	4.68	5.62	7.62	10.62	7.30	6.25	5.33	4.54
MnO	0.17	0.17	0.21	0.17	0.20	0.13	0.16	0.16	0.19	0.10	0.16	0.12	0.06
MgO	6.06	6.43	6.12	7.42	8.61	11.60	11.20	10.10	6.92	7.34	3.82	3.26	2.66
CaO	10.80	9.16	8.14	9.54	12.13	18.20	15.50	14.00	12.90	3.25	4.59	2.17	2.58
Na_2O	3.39	4.56	4.68	2.25	1.65	0.71	0.84	1.46	1.45	5.53	4.42	5.31	7.95
K_2O	0.68	0.53	1.04	1.79	0.35	0.03	0.07	0.08	0.07	0.56	1.42	2.20	0.19
P_2O_5	0.05	0.11	0.18	0.10	0.10	0.00	0.00	0.00	0.01	0.11	0.08	0.08	0.07
LOI	5.16	2.87	1.85	3.63	2.34	1.53	1.16	2.68	1.00	2.02	1.60	1.16	3.20
Trace elements (ppm)													
Ni	66	85	110	97	107	199	116	125	90	76	50	67	24
Cr	148	324	160	145	164	672	551	385	117	300	52	44	31
V	330	389	495			89	123	158	278	90	86	99	80
Sc						37	41	55	63	17	17	14	11
Zr	71	91	135			14	3	3	24	84	106	97	90
Nb	<1	2	3			2	<1	<1	<1				
Y	31	38	48			2	1	5	6	17	13	20	
Sr	188	153	103			47	43	48	37	232	259	199	163
Rb	17	10	24			2	<1	3	2	86	25	28	11
Ba						33	35	50	400	132		642	151

Table 3.2: Whole-rock analyses for Cambrian greenstones from the Stawell and Bendigo zones. 1–3: Magdala Volcanics: tholeiitic metabasalts from underground drilling, Stawell old mine (A. J. Crawford, unpublished). 4–5: Pitfield Volcanics: Greenstones from mine dumps at Pitfield (R. W. R. Ramsay, unpublished). 6–9: Ceres Metagabbro near Geelong (V. J. Morand, unpublished). 10–13: Lazy Bar Andesite: low-Ti andesite lavas associated with boninite in the central section of the Heathcote Greenstone Belt (A. J. Crawford, unpublished).

reported from the Magdala Volcanics at Stawell (Wilson *et al.,* 1992) probably reflects incorporation of old crustal lead. A two-end-member mixing age of 518±52 Ma better represents the age at which the rocks were extruded (Crawford, 1994; D. Foster, personal communication, 1999).

Pitfield Volcanics

Metamorphosed volcanic and intrusive rocks, including ultramafic rocks of tholeiitic affinity, occur as small fault-bounded slices along the Avoca Fault Zone from south of Pitfield, north to Burkes Flat (Fig. 3.1). They are known as the Pitfield Volcanics (Taylor *et al.,* 1996) and their age is assumed to be Cambrian on the basis of their similarity to Cambrian tholeiitic rocks at Heathcote. Rock types are mainly foliated to massive basalt and minor gabbro to dolerite, with green- schist facies assemblages. In the Pitfield area two mine dumps have yielded some serpentinised peridotite. The rocks do not outcrop, and have only been described from drill core and mine dump float, which includes some rounded cobbles from palaeodrainages (Ramsay *et al.,* 1996; Morand *et al.,* 1995).

The least-deformed basalts have relic igneous textures typical of submarine lava flows, varying from aphyric to mildly plagioclase- or clinopyroxene-phyric. Chlorite-rich layers with amygdules are interpreted as former glassy selvages that have been deformed and chloritised. Hyaloclastite occurs in drillcore from Burkes Flat. A few samples of medium-grained gabbro were probably high-level sills or dykes. Peridotites have cumulate textures, with original euhedral cumulus olivine now fibrous serpentine, set in a fine groundmass of serpentine and chlorite which may include unaltered intercumulus clinopyroxene. The most deformed rocks are intensely foliated greenschists with the metamorphic assemblage albite–chlorite–titanite–muscovite–carbonate–quartz±actinolite±epidote. Sulphides, dominantly pyrite, are also present in many samples.

Compositions of the Pitfield rocks (Table 3.2) are essentially the same as the tholeiites of the Heathcote and Mount Wellington Belts (Crawford & Keays, 1987; Ramsay *et al.*, 1996). They have MgO 7.4–8.6%, moderate TiO_2 (0.98–1.18) and immobile elemental ratios Ti/Zr, Zr/Y and Ti/V (av. 80, 3.3 and 23 respectively), similar to average mid-ocean ridge tholeiites at 7% MgO (Ti/Zr=110, Zr/Y=2.8 and Ti/V=22). The tholeiitic association of basalts and subvolcanic dolerite–gabbro sills is typical of low-K tholeiites, with geochemical signatures similar to depleted mid-ocean ridge basalts erupted in backarc basin-type settings (Crawford, 1988; Ramsay *et al.,* 1996).

St Arnaud Group

Overlying the tholeiitic basalts of the Stawell Zone is a pile of largely unfossiliferous, marine quartz–mica turbidites and occasional black shales known as the St Arnaud Group. Detrital zircon and mica populations show that the turbidites of the western Lachlan Fold Belt were derived from rocks uplifted during the Delamerian Orogeny at about the end of the Cambrian (Turner *et al.,* 1993). The only places where the base of the St Arnaud Group may be exposed are in the vicinity of Stawell, where quartz-rich turbidites appear to lie conformably on the Magdala Volcanics. With the exception of the complex rocks of the Moornambool Metamorphic Complex, the St Arnaud Group forms the bedrock across the entire Stawell Zone.

The St Arnaud Group is subdivided into three formations: the Warrak, Beaufort and Pyrenees formations on the basis of differences in sand to silt ratio, bed thickness, composition, and facies characteristics (Cayley & McDonald, 1995). Contacts between the Warrak Formation and the Beaufort and Pyrenees formations are faulted and the stratigraphic relationship are not known. The Warrak Formation is the oldest unit of the group, overlying Magdala Volcanics in Stawell Mine boreholes. It can be distinguished from its correlative in the Delamerian Fold Belt to the west, the Glenthompson Sandstone, because the latter has a significant component of coarse detrital mica flakes and is generally more texturally and compositionally immature. A Late Cambrian age is inferred for the Warrak Formation from the regional geological setting. It has a minimum thickness of 2–2.5 km.

The Beaufort Formation (Fig. 3.11) is at least 1–1.5 km thick and is relatively rich in siltstone, with sandstone:mudstone ratios ranging from 0.5:1 to 1:1. Structural relationships suggest that the Pyrenees Formation conformably overlies the Beaufort Formation. The monotonous lithology and tight folding of the Pyrenees Formation make it impossible to determine the thickness; about 2.5 km is a rough estimate. This formation is distinguished by a much higher sandstone content and by much greater bed thickness than other formations in the group.

Heathcote Greenstone Belt

The meridional-trending Heathcote Greenstone Belt provides one of the best-exposed sequences of Cambrian rocks in Victoria. It has been divided into three geologically distinct segments (Crawford, 1988):

1. A structurally simple southern segment south of the Cobaw Batholith that contains tholeiitic basalts, minor dolerite sills and sediments (VandenBerg, 1992);

2. A structurally complex central segment around Heathcote dominated by andesitic and boninitic volcanics and hemipelagic sediments and with intercalated fault slices of Ordovician turbidites and black shale;

3. A northern segment dominated by tholeiitic dolerite, basalt and hemipelagic sediment, underlain by minor boninite and volcanic sediment.

Revision of the Cambrian stratigraphy has been made possible by a combination of geochemical work (Crawford, 1982, Crawford *et al.,* 1984), new detailed mapping (Gray & Willman, 1991a; VandenBerg, 1992; Edwards *et al.,* 1998; Spaggiari *et al.,* 2002b) and acquisition of new geophysical data. A new lithostratigraphic group called the Heathcote Volcanics has been defined (Edwards *et al.,* 1998). It incorporates three newly defined formations, the Sheoak Gully Boninite, Lazy Bar Andesite and the expanded Mount William Metabasalt. These formations are overlain by the Knowsley East Shale (Thomas & Singleton, 1956) and Goldie Chert (VandenBerg, 1992). The base of the sequence is always the Mount William Fault. The top of the Cambrian sequence is faulted against Ordovician sediments in the central segment, and is also possibly faulted in the northern segment. In the southern segment, the Cambrian sequence passes conformably up into the overlying Ordovician Castlemaine Group of the Bendigo Zone.

The Sheoak Gully Boninite and the Lazy Bar Andesite outcrop in a series of fault slices in the central segment of the Heathcote Greenstone Belt. There are minor outcrops of Sheoak Gully Boninite in the northern segment. Boninite is also interpreted to occur under cover along the eastern margin of the northern segment and to the north of the outcropping belt, extending beyond the New South Wales border (Fig. 3.12). The remainder of the outcrops of volcanics in the northern and central segments and in the entire southern segment are composed of Mount William Metabasalt. This is conformably overlain by Middle and Upper Cambrian sediments, but relationships with the other volcanics are

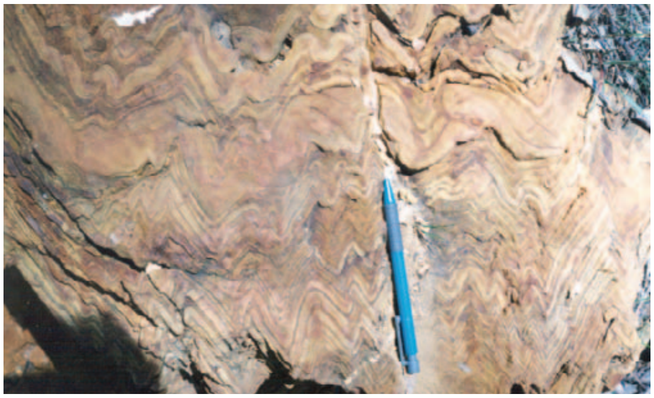


Fig. 3.11: Thin-bedded, silty turbidites typical of the Beaufort Formation, Freemans Road Creek, east of St Arnaud. The angular folds occur in the hanging wall of the west-dipping St Arnaud Fault. Photograph by J. Krokowski de Vickerod.

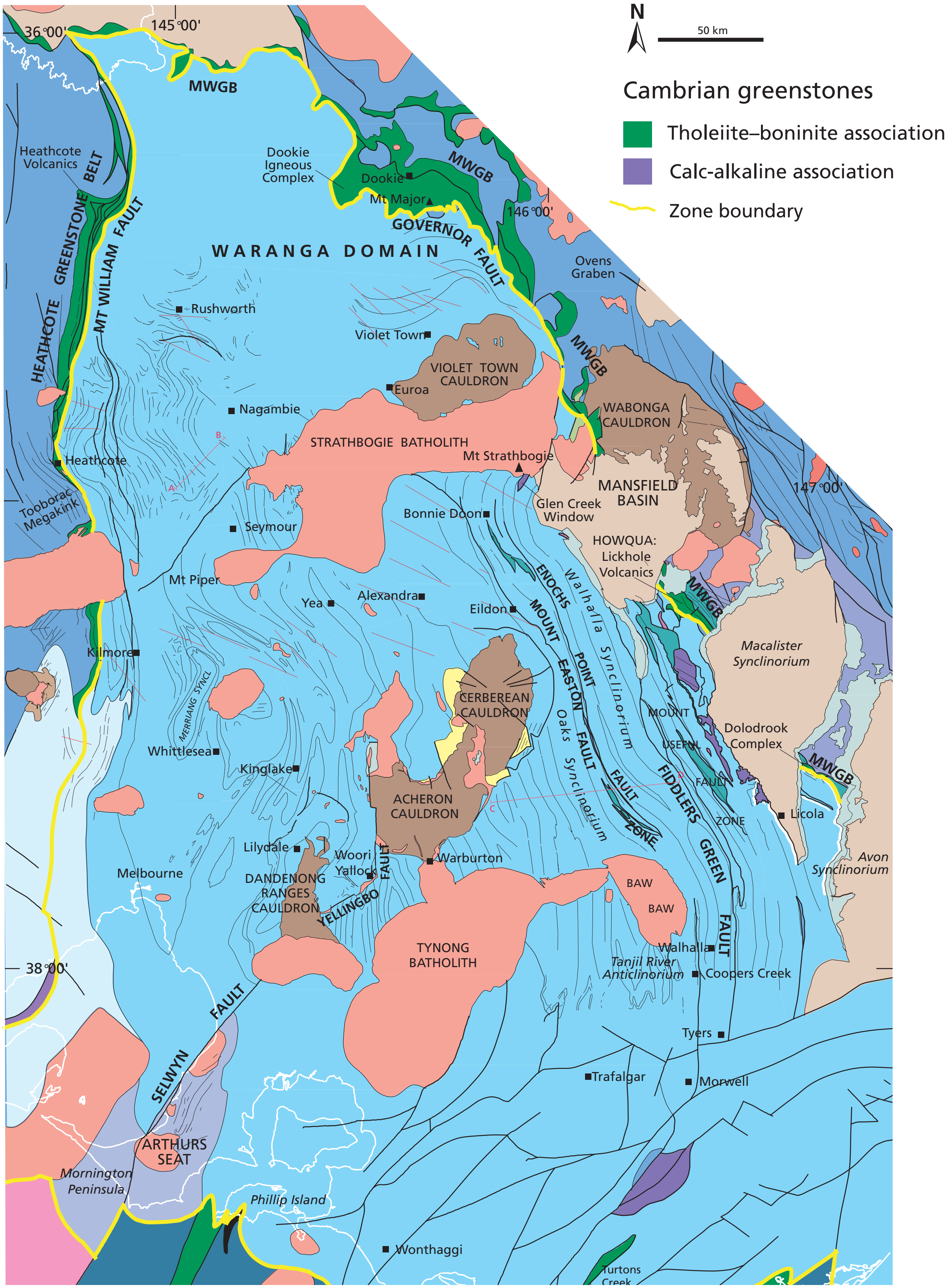


Fig. 3.12: Regional geological setting of central Victoria showing distribution of Cambrian greenstones in the Heathcote and Mount Wellington Greenstone belts, with subcrop along the Murray River inferred from aeromagnetic data (modified from VandenBerg *et al.*, 2000). MWGB = Mt Wellington Greenstone Belt.

invariably faulted. The northern segment of the Heathcote Greenstone Belt is a 2–4.5 km wide belt that extends northwards from Mount Camel under Murray Basin cover to just beyond the New South Wales border (Fig. 3.12). Rock exposure is restricted to a 2-km wide band along the western margin of the belt between Mount Camel and Rochester, but its complete distribution is evident in the aeromagnetic data (Fig. 3.5). The outcropping sequence consists of north-trending, concordant packages of Cambrian volcanics and pelagic sediments. Large-scale structures are interpreted from aeromagnetics and include duplexing in the easternmost belt of rocks (possibly volcanics) and an antiformal stack just north of Rochester. It is likely that the upper (?)volcanic package is similarly deformed and separated from the overlying Castlemaine Group by a fault.

Spaggiari *et al.* (2002b) recognised the central segment of the Heathcote Greenstone Belt as a fault-bounded mélange zone. It includes an intensely deformed belt of rocks some 7 km long that consists of fault slices of both Ordovician and Cambrian rocks. The internal deformation of the fault slices varies from moderately deformed in the Ordovician rocks with well-preserved sedimentary structures and fossils, to variably foliated Cambrian volcanics with patches of preserved igneous textures, and highly contorted Cambrian sediments (Gray & Willman, 1991a). Small (to 2 m long) blocks of blue schist contain the blue amphibole winchite and record metamorphic conditions of <450ºC and between about 500 and 700 Mpa (Spaggiari *et al.,* 1998, 2002b).

The southern segment is much less deformed; no décollement exists here between the Cambrian rocks and the overlying turbidites, so that a relatively undisturbed and conformable sequence is preserved. Greenstones, including lower dolerites and overlying basalts and intercalated pelagic sediments of the Mount William Metabasalt, are conformably overlain by the Knowsley East Shale (formerly called Monegeetta Shale, 200–500 m thick). These shales are, in turn, conformably overlain by the Goldie Chert, consisting of 190–290 m of chert and siliceous siltstone (VandenBerg, 1992).

Sheoak Gully Boninite

The Sheoak Gully Boninite consists of boninite lava and minor boninitic volcaniclastics and rhyolitic lava. The largest outcrop area occupies two of the largest fault slices in the duplex system of the central segment of the Heathcote Greenstone Belt. All the observed contacts of the Sheoak Gully Boninite are faults. Boninitic lava dominates the formation, mostly as massive coherent lava flows and minor pillow lavas. These lavas are a distinctive fine-grained blue–green rock, and igneous textures are best preserved at Sheoak Gully and Cornella East; elsewhere, the boninites are extensively altered and schistose. North of Ladys Pass the lavas have been contact-metamorphosed by the Crosbie Granite.

The boninites are high-Mg lavas containing up to 30% phenocrysts of low-Ca pyroxene which are mainly pseudomorphed by pale green chlorite (Crawford, 1982, 1984). The matrix of recrystallised glass contains euhedral, low- to high-Ca pyroxene as skeletal grains, dendritic aggregates and spherulites (Fig. 3.14a,b). Minor euhedral chromite grains occur in the matrix and as inclusions in the pyroxene phenocrysts. Ubiquitous quench textures indicate that the boninites probably formed thin lava flows.

A small outcrop of rhyolite lava occurs in Sheoak Gully (Edwards *et al.,* 1998). Patchy flow banding is defined by stretched vesicles up to 5 cm long which are filled by chalcedony and calcite. The formerly glassy lava displays well-developed devitrification textures, and spherulites are visible in hand specimen. The rhyolite contains embayed quartz and minor sericitised plagioclase phenocrysts in a commonly perlitically fractured and spherulitic groundmass consisting of devitrified glass, quartz and feldspar with minor opaque minerals. Calcite and sericite are common alteration minerals. The rhyolite lava has been partially quench-fragmented to form hyaloclastite and rhyolite breccia. The hyaloclastite contains fragments (to 5 cm) of rhyolite lava with jigsaw-fit textures, in a groundmass of devitrified glass. The rhyolite breccia contains large (to 10 cm), poorly sorted, angular to subrounded fragments of rhyolite in a matrix of chlorite-altered perlite. Margins of rhyolite fragments are marked by a thin zone of intense devitrification. The rhyolite in the Sheoak Gully Boninite is assumed to be a differentiate of the boninitic magma, but further geochemical work on the rhyolite is required to determine its petrogenesis.

Sediments interbedded within the Heathcote Volcanics near Heathcote have yielded a single Early Cambrian dolichometopiid trilobite (P.A. Jell, personal communication in VandenBerg, 1992). The exact location of this trilobite discovery is unknown.

Lazy Bar Andesite

The Lazy Bar Andesite is a generally poorly exposed and weathered sequence of andesitic lava, andesitic volcaniclastic sandstone, and minor vitric ash-rich fine-grained volcaniclastics (Nicholls, 1965; Crawford, 1982; Crawford *et al.,* 1984; Crawford & Cameron, 1985). The formation is restricted to the central segment of the Heathcote Greenstone Belt in several fault slices between the Cobaw Batholith and the Crosbie Granite. The andesite structurally overlies the Sheoak Gully Boninite with a concordant contact that is probably a fault.

The Lazy Bar Andesite lavas are generally fine-grained and non-vesicular. Where fresh, they contain euhedral phenocrysts of plagioclase, chloritised orthopyroxene and augite in a glassy groundmass, with grains of plagioclase, pyroxene and Fe–Ti oxide. Plagioclase is albitised and commonly overprinted by chlorite and epidote. The chartacteristic lower greenschist metamorphic assemblage is actinolite–chlorite–albite–epidote–quartz– leucoxene. Near the Heathcote Fault the andesite is altered to talc-actinolite schist.

The andesitic lavas were erupted into a marine environment. Sedimentary features indicate that the andesitic volcaniclastic sandstones were deposited by low-concentration turbidity currents, probably generated on the steep slopes of the growing andesitic volcanic pile. Ash-rich beds, including possible pumice, are interbedded with the lavas and indicate that the andesitic volcanic pile grew to a shallow enough depth for pyroclastic fragmentation to occur.

Geochemical and isotopic data suggest that these boninitic andesites were not derived by crystal fractionation from the underlying more mafic boninites, but rather, appear to have been derived from the same shallow refractory mantle source as the boninites by lower degrees of partial melting (Nelson *et al.,* 1984; Crawford & Cameron, 1985). There is little doubt that the Lazy Bar Andesite and Sheoak Gully Boninite are closely related temporally, and were produced in the same tectono-magmatic setting.

Mount William Metabasalt

The Mount William Metabasalt encompasses the tholeiitic volcanic sequence and associated sediments that outcrop along virtually the entire length of the northern and southern segments of the Heathcote Greenstone Belt. The formation is composed predominantly of thick sills of dolerite and basalt flows, some pillowed (Fig. 3.13), with minor bands of siliceous sediment (chert and jasper). It is overlain by the Knowsley East Shale in both the northern and southern segments. In the southern segment this contact is considered to be sharp, concordant and possibly conformable (VandenBerg, 1992). The Mount William Metabasalt is at least 2.5 km thick in the northern segment and over 1.6 km in the southern segment. The lower contact of the formation with the Sheoak Gully Boninite is faulted by the Corop Fault along the northern segment of the greenstone belt. South of Mount Camel, the lower contact with the Sheoak Gully Boninite is also faulted. However, as evolved tholeiitic dolerite dykes are known to cut the boninitic lavas east of Toolleen, the tholeiites clearly post-date the boninites and presumably overlie them.

The dolerite is a medium- to coarse-grained, green–black rock consisting of sparse phenocrysts of albitised plagioclase, euhedral to subhedral crystals of augite and minor opaques. Ophitic textures are common, and low greenschist metamorphic assemblages (chlorite–actinolite–albite–sericite-epidote) are widespread. Basalts are dominantly massive, often aphyric lava with minor pillow lavas. Sparse euhedral phenocrysts of albitised plagioclase and augite may be present (Crawford & Keays, 1987) (Fig. 3.14c). Metamorphic assemblages vary from prehnite–pumpellyite facies, with augite and anorthitic plagioclase preserved in places, to greenschist facies assemblages with actinolite after augite, albitised plagioclase, common epidote and chlorite, and leucoxene after the former Fe–Ti oxides.

Fig. 3.13: Pillows in Mount William Metabasalt, exposed in the Lake Cooper quarry, north end of Heathcote Volcanic Belt. Photograph by R. Cayley.

These basalts and dolerites are low-K tholeiites with flat REE patterns, and trace element signatures like MORB-type basalts generated in extensional zones (backarc basins and forearc extensional zones) above modern West Pacific-type subduction zones (Crawford & Keays, 1987),

Knowsley East Shale

Overlying the Mount William Metabasalt is the Knowsley East Shale, which includes the Middle Cambrian black shale and volcaniclastic units in all segments of the Heathcote Greenstone Belt. The formation includes the entire Knowsley East beds and a large part of the Goldie Beds of Thomas (1956), as well as the Goldie Shale sediments in Trilobite Gully described by Wilkinson (1977). It does not include the Goldie Chert at Lancefield in the southern segment. Other components of the Knowsley East Shale include minor interbedded chert, mafic lithic sandstone, polymictic conglomerate, monomictic chert breccia, and ash.

The formation is exposed along the western margin of the northern segment and as fault slices within the central segment, between Tooborac and Heathcote. It also commonly occurs as fault-bounded blocks which are juxtaposed against various other units of the Heathcote Volcanics and the Castlemaine Group. In the southern segment at Lancefield, the formation is conformably overlain by the Goldie Chert (VandenBerg & Stewart, 1992), but in the northern and central segments the boundaries are not so clear, and are probably faulted. A felsic airfall tuff in the Knowsley East Shale at Lancefield has a population of small equant zircons with a Middle Cambrian age (503±8 Ma). This age matches well zircon ages for the Mount Stavely Volcanic Complex (Stuart-Smith & Black, 1994; Crawford *et al.,* 1996a,b) and the compositionally similar Mount Read Volcanics in western Tasmania, both of which have been interpreted as post-collisional magmatic suites (Crawford *et al.,* 1992, 1996a,b). Although no similar post-collisional lavas are present in the Heathcote Greenstone Belt, the existence of zircons of this age suggests close proximity of the Heathcote Belt to the post-collisional magmatism now exposed in the Stavely Greenstone Belt and in erosion windows through the Melbourne Zone rocks further east at Jamieson and Licola.

At Heathcote, the basal portion of the Knowsley East Shale contains a spectacular upward-thinning package of graded sandstone beds deposited as turbidites or grain flows and composed of pyroxene crystals with minor feldspar. Thick beds of polymictic conglomerate at higher levels contain clasts of black shale, chert, jasper and mafic lava in a sandstone matrix.

The shales and mudstones of the formation accumulated as hemipelagic mud but the coarser clastic rocks were deposited by various types of gravity flows. The coarsest rocks all occur at the base of the formation and consist mainly of mass flows derived from a variety of igneous and sedimentary rocks. Such material ceased to be a significant component for the remainder of the formation, probably because the source region of the volcanic edifice became buried by pelagic sediments. Translational slide deposits, mainly chert breccias, show that the pelagic sediments were deposited on a slope which was sufficiently steep to be unstable.

Goldie Chert

The Goldie Chert is a wholly pelagic sediment, deposited below wave-base in the marine environment. In the southern segment the Goldie Chert contains abundant phyllocarid crustacea and a single conodont, either *Cordylodus angulatus* or *C. rotundatus*, indicative of Datsonian age (I. Stewart, in VandenBerg, 1992), very close to the Cambrian–Ordovician boundary. The depositional process was by grain-by-grain or aggregate settling of sediment through the water column. The presence of Goldie Chert in the central segment is uncertain (VandenBerg, 1992). Intensely deformed siliceous shale and chert at Ladys Pass in the central segment is tentatively assigned to the Goldie Chert, but distinguishing these sediments from the silicified shale of the Knowsley East Shale is difficult. At Ladys Pass, these rocks form a lozenge-shaped fault slice surrounded by Lazy Bar Andesite. The chert is absent from the northern segment.

Regional synthesis

The boninite–tholeiite association in the Heathcote Greenstone Belt is best matched on the modern Earth by extensional forearc regions of intra-oceanic arcs (e.g. Bonin–Mariana, North Tonga). Taking into account the 600-Ma, east-facing passive margin represented by the Glenelg and Grampians–Stavely zones, the Delamerian Orogeny may have involved collision of this Cambrian forearc with the leading edge of the passive margin. Thus, during the Cambrian, the Stawell and Bendigo zones appear to have been a deep marine ocean basin just outboard of the Delamerian Orogen as it was accreted to the Australian margin. In the west, the terrigenous turbidite sequences of the St Arnaud Group are interpreted, on the basis of provenance, isotope geochemistry and inherited zircon population, to be the detritus shed from the newly accreted and uplifted Delamerian Orogeny. The Cambrian chert and shale sequences at Heathcote, further to the east, appear to be time-equivalent but more distal deposits, accumulating here until the start of the Ordovician. By this time the turbidite fan had prograded east to this point, as recorded by the commencement of Castlemaine Group deposition.

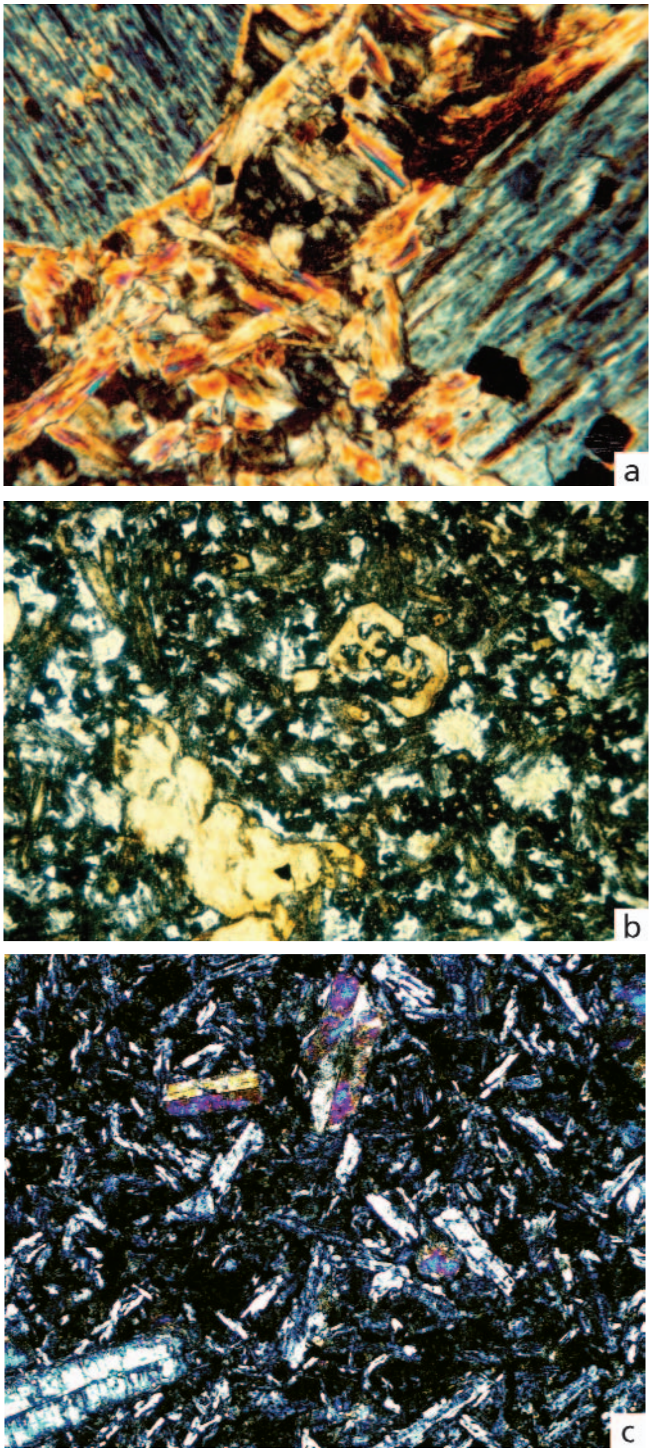


Fig. 3.14: Photomicrographs of typical textures in the Heathcote Volcanics (a,b from Sheoak Gully Boninite). (a) Low-Ca boninite 26278 (north Sheoak Gully) showing chlorite pseudomorphs after clinoenstatite phenocrysts in a groundmass charged with actinolite-altered pyroxene microlites and altered glass. Note the euhedral chromite inclusions in the altered clinoenstatite phenocrysts (bottom left). (width of field ~4mm). (b) Low-Ca boninite 26279 (Sheoak Gully) showing typical chlorite-altered quenched orthopyroxene microphenocrysts with cruciform growths extending into formerly glassy hollow centres of crystals, and a groundmass charged with pyroxene microlites set in glass that has altered to chlorite and quartz. (width of field ~4mm). (c) Typical interior of thick tholeiitic basalt flow, showing fresh clinopyroxene (augite) plates, partly fresh and partly sericite-albite-altered plagioclase laths often partially included in the augite, common fine-grained Fe-Ti oxides and interstitial chlorite. From old rail cutting at Kilmore Pass, Mt William Metabasalt. All images are about 5 mm across and are taken with crossed polars.

3.3.2 Melbourne Zone

The Melbourne Zone consists of deformed Ordovician to Devonian sedimentary rocks. Along the eastern margin of the zone, erosion has carved windows through the highly sheared base of this metasedimentary sequence to expose a basement of Cambrian volcanic rocks (Gray, 1995; VandenBerg *et al.,* 1995).

Jamieson–Licola Volcanics

This belt consists of four major exposures or 'windows' with irregular outlines within the Mt Useful Fault Zone. New airborne geophysical data and mapping by the Victorian Geological Survey (VandenBerg *et al.*, 1995), along with contributions from university researchers (Hendrickx, 1993; Cherry, 1999), have significantly revised the shape of the greenstone exposures and clarified their internal stratigraphy. The base of these volcanic rocks, which have been referred to informally as the 'Barkly River Greenstone Belt' (Turner, 1996; Cherry, 1995), is not exposed. Until recently these exposures were interpreted as fault slivers incorporated into the basal parts of the Mount Wellington Fault Zone (Fergusson *et al.*, 1986; Gray, 1995). A more recent appraisal re-interprets them as erosional windows through the highly sheared base of the Mount Useful Fault Zone into the underlying Selwyn Block (VandenBerg *et al.*, 1995, 2000) (Fig. 3.15). The Jamieson Window is defined by aeromagnetic data and the response appears to continue southwards, linking with the magnetic highs of the Whisky Knob Window, and implying a subsurface link between the two windows. South of the Whisky Knob Window are the Fullarton Spur and Licola windows (Fig. 3.12). These two southern windows have much more subdued magnetic responses compared with the two northern windows, implying different rock packages. Much further to the north of all these occurrences is a poorly known exposure at Glen Creek (Fig. 3.12).

Jamieson and Whisky Knob windows

The Jamieson Window has been mapped in detail by Hendrickx (1993) and summarised by VandenBerg *et al.,* (1995), who subdivided the stratigraphy within the window into a number of formations. These include the Brissces Hut Andesite, the Warrambat Andesite Breccia, Wrens Flat Andesite, Lakelands Flat Andesite Breccia, Hardwicke Creek Rhyolite, and the Handford Creek Formation. Subsequent mapping by Cherry (1999), supplemented by geochemical data, suggested that the andesite and andesite breccia formations erected by VandenBerg *et al.,* (1995) might better be considered as a single stratigraphic unit, and that bedding mostly trends almost east–west, rather than NW-trending as reported earlier. Dips are mainly to the south. Cherry (1999) estimated that coherent lavas form less than 50% of the volcanic–volcaniclastic package. Coarse sandstone and mass flow breccia are abundant, with siltstone common. Lavas are mainly plagioclase+augite-phyric andesite with good textural preservation; monomictic andesite lava breccias are common. Metamorphic assemblages are low greenschist facies. At several localities, occasional clasts in polymictic breccia show a tectonic foliation. High-level intrusive andesitic rocks probably represent thin sills and dykes but are difficult to distinguish from the andesitic lavas. Geochemical data indicate that the intrusive rocks and andesitic lavas are comagmatic. It is very likely that the thick andesitic pile east of the Hardwicke Creek Rhyolite is fault-repeated, possibly several times.

A felsic sequence constituting the Hardwicke Creek Rhyolite overlies the andesitic sequence and occupies much of the southern third of the Jamieson Window. Cherry (1999) reported rhyolitic and dacitic lava and lava breccia, many of which carry quartz phenocrysts, and common fine- to medium-grained thinly bedded sandstone. Rare andesites with hornblende and plagioclase are probably lavas. The Hardwicke Creek Rhyolite has an interpreted true thickness of about 1100 m, but its base is marked by a major shear zone several tens of metres wide. Its upper contact appears to be faulted against the overlying Handford Creek Formation (Hendrickx, 1993), although Cherry (1999) suggested that this faulting may only be local and that the contact may be conformable over much of its length.

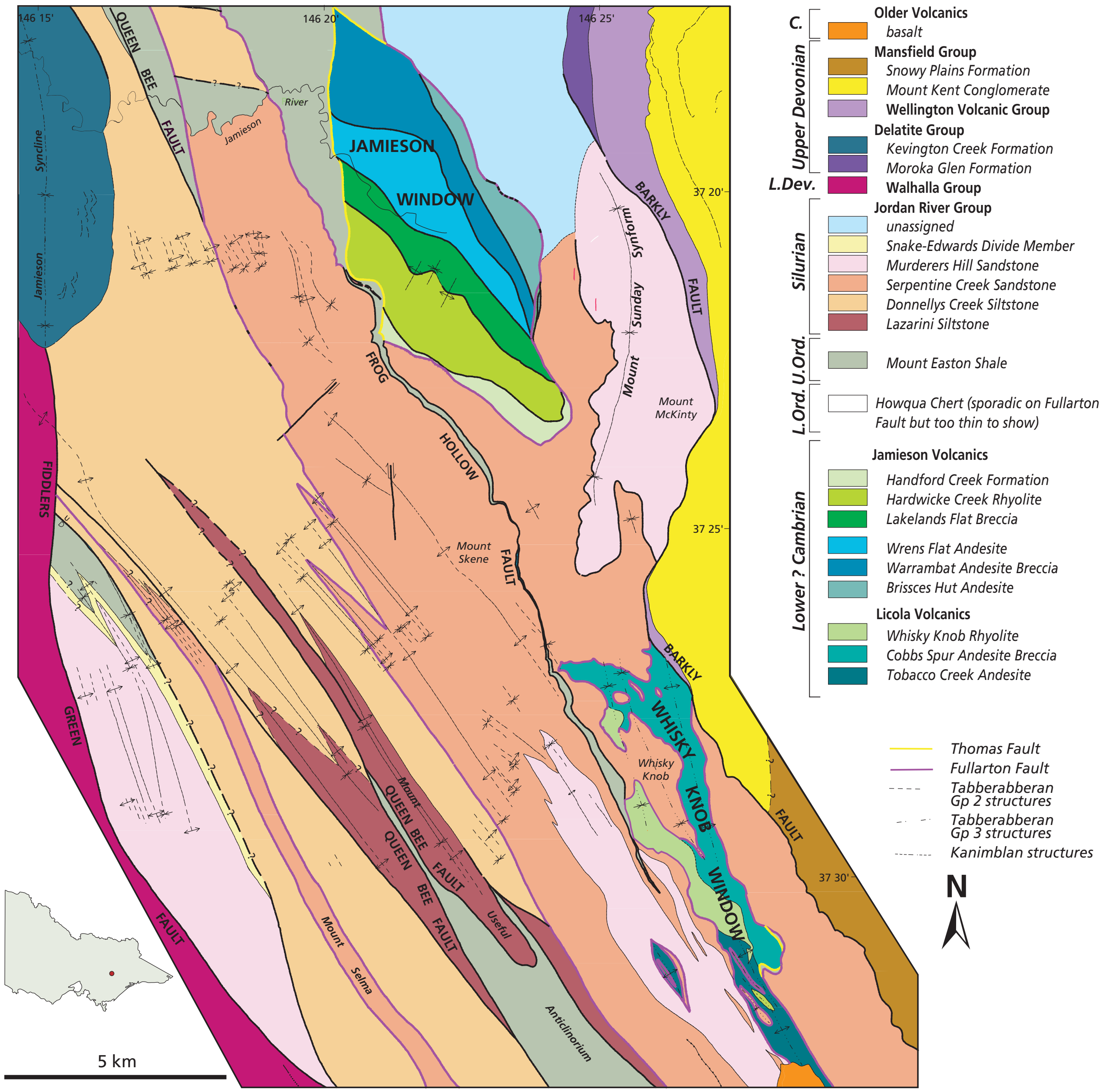


Fig. 3.15: Distribution of fault slices of Licola and Jamieson volcanics in the Mt Useful Fault Zone, eastern part of the Melbourne Zone (modified from VandenBerg *et al.,* 2000).

The Handford Creek Formation forms the top few hundred metres of the Cambrian stratigraphy in the Jamieson Window. It contains mainly volcaniclastic sedimentary rocks, with no lavas or lava breccias. Massive poorly sorted sandstones carry common volcanic quartz detritus and volcanic lithic clasts. Granule- to cobble-sized conglomerate beds 1–2 m thick occur within the finer grained sequence. Clasts include cherty rocks, felsic lavas and possible pumice fragments.

Geochemical data (Crawford, 1982, 1988; Cherry, 1999) show that the Jamieson rocks are medium-K calc-alkaline andesites (Table 3.3). The limited range of compositions among the analysed samples collected from a wide area of the northern part of the Jamieson Window suggests that a single lithostratigraphic unit is represented among the extensive andesitic lavas and lava breccias. This supports the suggestion by Cherry (1999) that the four formations erected by VandenBerg *et al.,* (1995) are better considered as a single lithostratigraphic unit, albeit probably repeated by thrust faulting. Further geochemical studies of the felsic volcanics are required to evaluate their significance.

The stratigraphy of the Whisky Knob Window is less well known, but rhyolites are common and are petrographically close to those in the southern section of the Jamieson Window; these remain to be studied geochemically.

Fullarton Spur and Licola windows

The Fullarton Spur and Licola windows are less well known, due largely to difficult access. The excellent exposure on the Jamieson-Licola Road in the Licola Window is of columnar high-K hornblende andesite (the Tobacco Creek Andesite). This andesite is strikingly similar petrographically and compositionally to the 500-Ma Anthony Road Andesites of the Mount Read Volcanics in western Tasmania (Crawford *et al.,* 1992, 1996a,b).

	1	2	3	4	5	6	7	8	9	10	11	12
	26295	**26296**	**26298**	**26299**	**26304**	**26309**	**26362**	**26363**	**26364**	**E12498**	**E12299**	**E12292**
SiO_2	57.70	55.60	60.30	66.20	60.60	62.20	50.20	49.30	50.60	49.20	50.50	48.20
TiO_2	0.48	0.46	0.41	0.42	0.48	0.46	2.18	1.52	1.73	0.86	0.98	0.28
Al_2O_3	14.20	14.90	12.80	12.60	15.00	14.70	13.70	14.10	13.10	14.70	14.90	16.00
FeO*	8.96	9.26	7.70	6.03	6.28	6.40	17.30	13.90	14.20	10.10	9.85	9.05
MnO	0.14	0.13	0.10	0.09	0.12	0.10	0.27	0.16	0.22	0.22	0.22	0.20
MgO	5.92	6.82	5.34	2.99	4.66	3.40	5.94	8.21	6.89	10.10	9.65	14.20
CaO	8.30	8.59	8.33	7.35	7.70	5.85	7.41	9.52	9.53	10.50	9.00	8.36
Na_2O	2.53	2.83	4.07	1.52	2.78	4.03	2.41	2.89	3.37	3.23	4.40	1.92
K_2O	1.64	1.19	0.82	2.65	2.18	2.60	0.47	0.29	0.27	0.25	0.23	1.47
P_2O_5	0.17	0.16	0.14	0.19	0.26	0.25	0.19	0.12	0.14	0.10	0.08	0.02
Loss	2.31	3.12	1.27	1.47	2.80	3.68	4.02	2.54	1.64	1.12	0.82	1.25
Trace elements in ppm												
Ni	25	16	27	21	53	42	27	64	61	206	120	344
Cr	133	159	173	119	167	150	47	117	74	546	377	2143
V	233	237	194	218	164	161	509	340	375	256	265	179
Sc	26	27	25	22	15	17	42	41	41	35	47	34
Zr	108	80	72	102	216	161	94	65	99	56	48	15
Y	20	18	17	22	28	25	41	24	37	20	19	17
Sr	831	443	172	830	2096	672	132	127	172			
Rb	47	34	26	20	68	87	16	7	10			
Ba	476	484	636	112	1431	1533	180	58	65			

Table 3.3: Whole rock analyses for Cambrian greenstones from the Mt Wellington Greenstone Belt. 1–4: Andesites from the Jamieson window (A. J. Crawford, unpublished). 5, 6: Andesites from the Licola Window (A. J. Crawford, unpublished). 7–9: Tholeiitic metabasalts from Cape Liptrap (Maitland Beach Volcanics) (A. J. Crawford, unpublished). 10–12: metabasalts and metadolerite (12) from Kitty Miller Bay, Phillip Island (Henry & Birch, 1992).

Glen Creek Window

At Glen Creek south of Mount Strathbogie (Fig. 3.12), andesitic greenstones of presumed Cambrian age (Sandl, 1989) may be correlatives of the Jamieson and Licola volcanics. No petrological or geochemical studies of these rocks have been undertaken. A lower unit consists of porphyritic andesite lavas, in part vesicular and with good flow textures. An overlying polymict breccia consists of shale and siltstone fragments, ultramafic material, basaltic andesite and possibly felsic volcanic clasts, impure quartzite and possible clasts of granitic origin (Sandl, 1989). Interbedded with the breccia are lithic quartz sandstones and carbonaceous mudstone and siltstone. The margins of the Glen Creek Window are presumed to be faulted.

Waratah Bay —— Maitland Beach Volcanics

At Waratah Bay (Fig. 3.1), Cambrian igneous rocks outcrop in a broad, NNE-striking horst block exposed between the Waratah and Bell Point faults. Although often referred to in the literature, they were not formally named until very recently (VandenBerg *et al.,* 2000; Cayley *et al.,* 2002). Rock units include metabasalt and interbedded pelagic sediments, metagabbro, and strongly altered olivine-rich ultramafic rocks. Most of the belt between the Waratah and Bell Point Fault zones consists of outcropping Maitland Beach Volcanics, with outcrops of the Corduroy Creek Gabbro and ultramafics confined to a relatively small area near Digger Island.

The Maitland Beach Volcanics were mapped by Lindner (1953) and Sandiford (1978). Crawford (1982) reported geochemical data (Table 3.3) for the tholeiitic basalts that form a major part of the exposed sequence. These basalts, which include rare pillowed flows and thin dykes, are either aphyric or sparsely augite+plagioclase-phyric, and show prehnite–pumpellyite or lowest greenschist facies metamorphic assemblages. On the basis of pronounced petrographic and geochemical similarities with Heathcote and Howqua tholeiitic basalts, the Maitland Beach Volcanics are interpreted to be Cambrian in age.

A major fault along the eastern margin of the Maitland Beach Volcanics is marked by zones of intense shearing and hydrothermal alteration, slices of dark recrystallised limestone, the coarse-grained Corduroy Creek Gabbro, and serpentinised and silicified peridotite. Previous authors have regarded the gabbro's age as either Devonian (e.g. Lindner, 1953) or Cambrian (Crawford, 1988). Its pre-Ordovician age is demonstrated by the unconformably overlying Cambrian Bear Gully Chert and Lancefieldian Digger Island Formation (Cayley *et al.,* 2002). Metamorphic hornblende in the Corduroy Creek Gabbro indicates significantly higher grade metamorphism than in the adjacent Maitland Beach Volcanics. Petrographic observations of the altered ultramafic rocks indicate that they were originally dunite. Iridium–osmium nuggets probably sourced from the serpentinised ultramafics have been found in Cainozoic placers up to 40 km away. The occurrence of these alloys invites comparison with the Cambrian mafic–ultramafic complexes of Tasmania, in which dunite is commonly associated with placer deposits of 'osmiridium' (Brown & Jenner, 1989). Geochemical data for the Maitland Beach Volcanics, together with their close association with gabbro and peridotite, are also reminiscent of the Cambrian ophiolite sequences in Tasmania (and Victoria), rather than the supracrustal Late Neoproterozoic rift tholeiite sequences in Tasmania, in which gabbro and peridotite are unknown.

The deformed Maitland Beach Volcanics and the Corduroy Creek Gabbro are both overlain by the Bear Gully Chert, a thin siliciclastic unit (Cayley *et al.,* 2002), and by the Early Ordovician Digger Island Formation, a shallow marine limestone unit (see Chapter 4). This contrasts markedly with other occurrences of presumed Cambrian greenstone in Victoria, for example along the northern and southern Heathcote Greenstone Belt, and at Howqua, where Late Cambrian – Early Ordovician deep-water pelagic sediments conformably follow the basaltic greenstones. The Bear Gully Chert is exposed immediately above the unconformity approximately 350 m north of Digger Island. It consists of fine-grained quartz and small lithic clasts, which form a matrix supporting larger angular to rounded deformed lithic clasts and occasional large rounded quartz pebbles. No clasts of the underlying meta-igneous rocks have been recorded. The unit is pyritic, with small pyrite crystals forming up to 20% of the rock. Although thin (<20 cm), this unit is crucial in fingerprinting the provenance of the deformation of the underlying meta-igneous rocks, as it indicates that uplifted, deformed continental siliciclastics were being shed onto the unconformity prior to the Lancefieldian, the age of the conformably overlying Digger Island Formation. The Bear Gully Chert can therefore be no younger than Lancefieldian, and a Late Cambrian age is most likely. This unit is a direct correlate in terms of age and lithology with the Owen Group of western Tasmania, which unconformably overlies the Tyennan unconformity there. The pre-Ordovician age of the unconformity at Waratah Bay therefore presents key evidence for the presence of the Tyennan or Delamerian Orogeny in central Victoria.

Phillip Island and Barrabool Hills

Greenstone exposed on the southern coast of Phillip Island (Henry & Birch, 1992; Bushby, 2001) (Fig. 3.16) appears to lie on a northward extrapolation of a major magnetic high which trends across Bass Strait to exposures of Neoproterozoic rift tholeiites and picrites on the southeast coast of King Island. Although tentatively correlated with the Neoproterozoic basaltic rocks of Tasmania and King Island by Cayley *et al.* (2002), the Phillip Island rocks include greenschist facies formerly glassy boninitic lavas, dolerite and cumulate ultramafic rocks with characteristic high-Cr chromites. Also, their geochemical signature matches better with the low-Ti dolerites and boninitic lavas and ultramafics of the Cambrian sequences exposed at Howqua and in western Tasmania, and is atypical of the Late Neoproterozoic rift tholeiite–picrite sequences of western Tasmania.

Fig. 3.16: Greenstone outcrops near Kitty Miller Bay, Phillip Island. Cambrian metavolcanics form the shore platform in the foreground and the small headland in the middle distance. Photograph by W. Birch.

Ceres Metagabbro

A small, little-known outcrop of metagabbro, the Ceres Gabbro, occurs in the Barrabool Hills near Geelong (Fig. 3.1). Small outcrops of the same metagabbro occur a few kilometres to the north at Dog Rocks. The metagabbro has no contacts with Palaeozoic rocks other than a few tiny granite intrusions of presumed Late Devonian age. The Ceres outcrops appear to form a thrust slice of gabbro that has been metamorphosed to amphibolite facies. The rock is mainly massive, coarse- to fine-grained, with sporadic subtle layering defined mainly by changes in grainsize (Morand, 1995; Cayley *et al.,* 2002). This layering is vertical and strikes northwest. Mg-rich diopside and some bytownite are the only igneous minerals preserved. Opaque minerals are rare, with Fe–Ti oxides occurring only in the most Fe-rich samples. The rock has equigranular gabbroic or rare subophitic textures; cumulate textures have not been observed. Analyses reveal that the rock is a tholeiitic gabbro (Table 3.2), with 47–52% SiO_2 contents on an anhydrous basis. Notable features are the moderately high MgO, low Na_2O and the very low TiO_2, K_2O and P_2O_5 contents.

The metamorphic assemblage is anorthite–calcic amphibole, with minor chlorite; most plagioclase is recrystallised into a fine-grained granoblastic aggregate of anorthite and most clinopyroxene is partly or completely replaced by Mg-rich amphibole. Many rocks are massive, but foliated to mylonitic samples with the same amphibolite facies metamorphic assemblage are common. This amphibolite facies metamorphism was accompanied by N–S compression, as indicated by a conjugate set of internal shear zones (Morand, 1995). This trend is at odds with the regional structures developed in the Bendigo Zone to the north (Morand, 1995). The metamorphism and deformation affecting the Ceres rocks appear to be regional, with the small post-tectonic granite plutons intruding the metagabbro appearing to have had little metamorphic effect.

Correlation of the Ceres Metagabbro with metabasic rocks of Neoproterozoic or Cambrian age in Victoria and Tasmania is not straightforward. Is it better correlated with the Late Neoproterozoic greenstone exposed on King Island and northwestern Tasmania, or with the tholeiitic gabbros such as those at Howqua? Geochemical data for the Ceres Metagabbro match well with the Howqua gabbros. In contrast, gabbroic rocks are limited to a few microgabbro sills and dykes in the Neoproterozoic packages of western Tasmania. Although it remains unproven, the Ceres Metagabbro is suggested to be a deep-derived, amphibolite-grade thrust-slice of Cambrian gabbros, such as those in the Heathcote and Mount Wellington Greenstone Belts.

Regional Synthesis

The calc-alkaline exposures along the eastern margin of the Melbourne Zone (Jamieson and Licola volcanics) show remarkable geochemical and petrographic similarity to the 500-Ma Mount Read Volcanics of western Tasmania. The interpretation that these exposures are erosion windows through the Melbourne Zone into an older basement — the Selwyn Block — which is the northern extension of Tasmanian crust, reinforces this suggestion (VandenBerg *et al.,* 2000). In this scenario, the Jamieson and Licola volcanics might be broadly interpreted as along-strike continuations of the Mount Read Volcanics.

3.4 Eastern Victoria (Tabberabbera Zone)

3.4.1 Mount Wellington Greenstone Belt

Introduction

The Mount Wellington Greenstone Belt occurs as a series of discontinuous fault slices in the hanging wall of the Governor Fault along the western margin of the Tabberabbera Zone (Fig. 3.12). Outcrop areas include Dookie, Tatong, Howqua and Dolodrook River, with aeromagnetic data suggesting more greenstone in the north under Murray Basin cover. The rocks are mainly the same boninite–tholeiite association as exposed along the Heathcote Greenstone Belt, and include the Dookie Igneous Complex at Dookie, the Lickhole Volcanic Group on the Howqua River and the Thiele Igneous Complex on the Wellington River (VandenBerg *et al.*, 2000).

Overlying the Lickhole Volcanic Group in the Howqua section is the Howqua Chert, approximately 500 m of mostly chert and siliceous shale with minor lithic sandstone, pebbly sandstone and chert conglomerate (Crawford, 1988). Only the uppermost portion contains useful fossils, which are basal Ordovician (Lancefieldian, La2) graptolites and conodonts at Howqua River. The only other recorded fossils are small inarticulate brachiopods. Volcaniclastics associated with the chert also occur in the hanging wall of the Wonnangatta Fault near Crooked River. The same rock unit overlies the Dookie Igneous Complex, where the chert consists of quartz and albite (Christie, 1978). The chert is highly pyritic when fresh. Minor components are graded volcaniclastic sandstone and conglomerate, and mudstone. Some conglomerates contain gabbroic detritus (Tickell, 1989).

Howqua Section

The best-studied volcanic sequence in this greenstone belt is the Lickhole Volcanic Group on the Howqua River (Crawford, 1982; Spaggiari *et al.,* 2002b). The basal part of the greenstone sequence is separated by a major fault zone from a 3-km-wide polydeformed mélange zone that includes blueschist blocks up to 5 m long, in which relict glaucophane and winchite suggest metamorphic conditions of <450ºC at 700–900 MPa (Spaggiari *et al.,* 2002b). The basal Mountain Chief Andesite is a thin (100–250 m) formation of andesitic volcaniclastics and mafic boninitic lava and hyaloclastite. Overlying this is the Sheepyard Flat Boninite, 1000–1500 m of ultramafic boninitic lava and volcanic breccia with remarkable textural variations and rare interbeds of finer volcaniclastics. Two thin flows of tholeiitic basalt occur close to the top of the pile. Above this is the Malcolm Creek Hyaloclastite, about 750 m of 5–10-m thick beds of tholeiitic hyaloclastite with occasional beds of pebbly grit and volcaniclastic sandstone. The hyaloclasts contain phenocrysts of fresh augite, sparse chloritised olivine and albitised plagioclase, and the volcaniclastics contain clasts of boninite, serpentinite and porphyritic andesite (Crawford, 1982). The uppermost volcanic unit is the thick Eagle Peaks Basalt, up to 1.5–2 km of pillowed and massive aphyric tholeiitic basalt with minor amounts of interflow and interpillow cherty sediment (Crawford & Keays, 1987).

Intercalated with the boninitic volcanics are an olivine pyroxenite sill, comagmatic with the Sheepyard Flat Boninite (Crawford, 1980), and sills and dykes of dolerite and gabbro comagmatic with the Eagles Peak Basalt (Crawford & Keays, 1987) (Fig. 3.17a,b). The largest sill is close to 500 m thick and shows significant compositional layering. The tholeiitic basalt and underlying boninite in the Mount Wellington Greenstone Belt are closely comparable in their geochemistry and petrography to those in the Heathcote Greenstone Belt (Crawford & Cameron, 1985; Crawford & Keays, 1987).

Dookie and Tatong sections

The sequence and rock relationships of the Cambrian rocks at Dookie are still not well understood. The exposed sequence is at least 1000 m thick, but bedding-parallel thrusts have probably been overlooked in previous studies of the area. There are three main rock types: metabasalt, gabbro and sediments. The basalts are all tholeiitic flows that appear to lie between two successions of similar sedimentary rocks. These sediments consist mainly of chert, which includes beds consisting of quartz and albite that were probably originally volcanic ash (Christie, 1978), and also rare detrital quartz grains and sponge spicules. Also present are black shale and siltstone, and sandstone and conglomerate with grains and clasts of basalt and gabbro. A thick sill of gabbro occurs in the 'lower', southern belt of sediments.

The outcrops at Tatong comprise a sequence of tholeiitic gabbro, dolerite and basalt,overlain by Howqua Chert and Pinnak Sandstone (McGoldrick, 1976; Crawford, 1988).

Dolodrook River Inlier

The Dolodrook River Inlier (Thiele Igneous Complex; Fig. 3.12) is an anticlinal structure cored by serpentinised ultramafic rocks (Fig. 3.18), and surrounded by incomplete rings of successively younger rocks ranging in age from early Middle Cambrian to Silurian (Teale, 1920; Duddy, 1974; Spaggiari, 2003). They form a curved elongate belt about 5 km long, trending northwest, parallel to the structural trend of the surrounding Silurian rocks. Ultramafic rocks have been shown (Crawford, 1982) to include cumulates from both boninitic and low-K tholeiitic magmas, and are almost certainly directly related to the boninite-tholeiite lava association exposed elsewhere in the Mount Wellington Greenstone Belt (e.g. at Howqua). A series of thinly bedded green sandstone and shale, with minor conglomerate, the Garvey Gully Formation, is at least 200 m thick and unconformably overlies the serpentinised ultramafics. Channel structures and rounding of coarse clasts suggest that the Garvey Gully Formation was deposited rapidly under shallow, moderate- to high-energy conditions. Intercalated in the upper part of the formation is the Dolodrook Limestone Member, a shelly algal pelletal limestone containing late Middle to early Late Cambrian trilobites. Fault-bounded, incomplete rings of Middle Ordovician sandstone and slate, and Late Ordovician dark shale and chert, surround the greenstone core.

Cambrian volcaniclastics have recently been discovered in the hanging wall of the Wonnangatta Fault, underlying the Howqua Chert, near Crooked River. The section is a few metres thick and consists of thin, graded beds of what were probably pyroxene sandstones, but are now talc–chlorite slate. The chert contains conodonts of the Datsonian (latest Cambrian) *C. proavus* Zone (VandenBerg *et al.,* 2000).

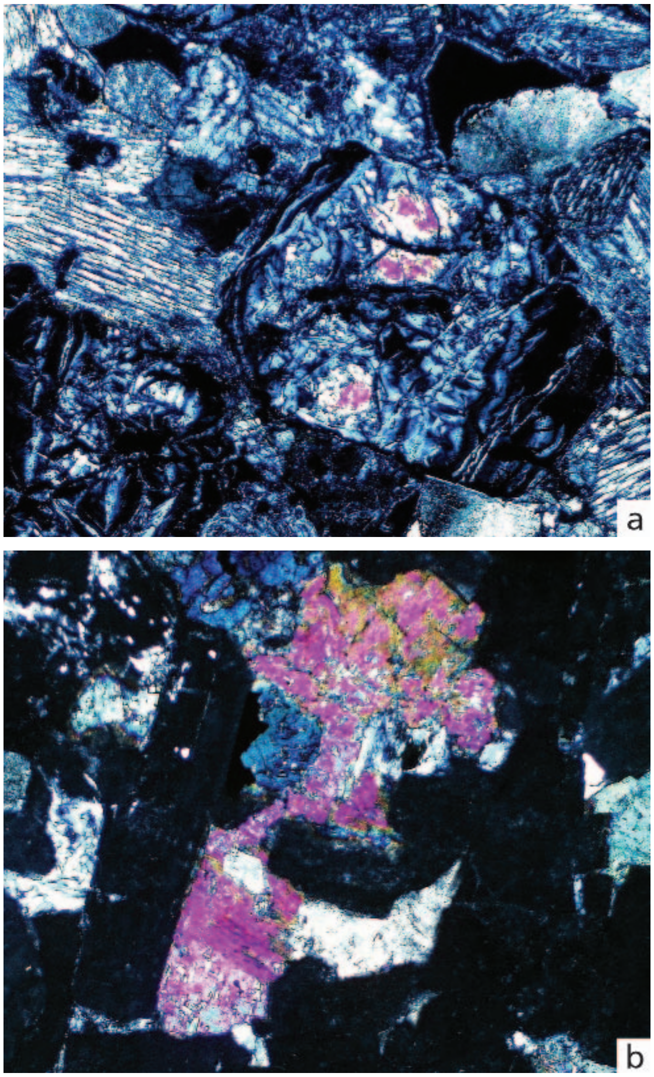


Fig. 3.17: Photomicrographs of rocks from the Lickhole Volcanics Group in the Mt Wellington Greenstone Belt. (a) Former ultramafic cumulate from boninitic magma, with crystals of clinoenstatite and rare olivine replaced by talc and tremolite-actinolite. From Cold Creek crossing on the Howqua Track. (b) Tholeiitic gabbro showing fresh clinopyroxene plates and smaller plagioclase crystals replaced by near-isotropic microcrystalline epidote. From large layered sill, Lower Howqua track. Both slides are c5 mm across and are taken with crossed polars.

Fig. 3.18: Serpentinised ultramafic host to chromite deposit, Dolodrook River, Thiele Igneous Complex (lens cap for scale). Photograph by W. Birch.

3.5 Summary

The Late Neoproterozoic and Cambrian rocks in Victoria differ from later Palaeozoic sequences by the predominance of volcanic and volcaniclastic rocks. It is convenient to consider the volcanic rocks within the framework of three broad tectono-magmatic associations that have relatively well-defined temporal constraints. These are:

Association 1

A latest Neoproterozoic rift–drift sequence around 590–600 Ma, dominated by rift tholeiites and some olivine-rich picritic lavas. This association is presently thought to be restricted to the Delamerian Fold Belt section of western Victoria, west of the Moyston Fault.

Association 2

An intra-oceanic arc association, which consists of allochthonous slices of boninitic lavas and their cumulate counterparts, and overlying backarc basin-type tholeiitic basalts, largely restricted to the Lachlan Fold Belt. Rocks of this association dominate the Magdala, Pitfield, Heathcote and Mount Wellington Greenstone Belts, but also occur as limited fault-bounded slices west of the Moyston Fault at Wartook and west of Moyston. Early Cambrian trilobites, and analogous sequences in western Tasmania dated at 514 Ma, suggest ages probably between 520 and 510 Ma. By analogy with better exposed sections of the same sequence in western Tasmania, this association was probably emplaced during the earliest phase of the Delamerian Orogeny (510–505 Ma). This occurred during collision of the forearc section of an intra-oceanic arc with the east-facing, rifted passive margin of the Delamerian Fold Belt (characterised by Association 1 and its sediment cover). The leading edge of this collision probably extended to western Victoria, probably as far as the Moyston Fault. More distal from the collision zone, in sections typified by the Heathcote and Howqua sections, there is no structural evidence for this collision, and the volcanics are conformably overlain by cherts and a deep marine sediment sequence that extends through to the end of the Ordovician. This is an important difference from the similar age rocks of western Tasmania, which were all deformed during the Cambrian Tyennan Orogeny. These are proximal to the collision zone, since Association 2 rocks sit immediately upon the Association 1 rift sequences, with amphibolitic mylonite soles recording west-directed emplacement of Association 2 (Berry & Crawford, 1988).

The best exposures of Association 2 in the Heathcote and Mount Wellington Greenstone Belts are basal duplexes of major east-directed thrust systems of probable Latest Ordovician to Early Silurian age (Gray & Foster, 1997).

Association 3

A post-collisional association is represented by diverse medium- to high-K calc-alkaline andesites and high-Mg andesites occurring mainly west of the Moyston Fault upon the recently deformed Delamerian Fold Belt rocks. It is also exposed along the eastern margin of the Melbourne Zone at Jamieson and Licola in what are probably erosion windows through to an older basement of Tasmanian affinity. The best-exposed sequence, in the Mount Stavely Belt, has been dated at about 500 Ma. In terms of age and composition, these volcanics resemble the Mount Read Volcanics in western Tasmania, for which a post-collisional setting can be demonstrated (Crawford & Berry, 1992).

Assembly of the basement elements across central and western Victoria is still poorly understood, but in broad terms, rock sequences record a mid-Cambrian collision between a west-facing intra-oceanic arc and an east-facing passive margin. Post-collisional extension at about 500 Ma produced widespread calc-alkaline magmatism, now largely preserved as fault-bounded slices.